Late Quaternary deep-ocean circulation

BRUCE H. CORLISS *Department of Geology, Duke University, Durham, North Carolina 27706*
DOUGLAS G. MARTINSON *Lamont-Doherty Geological Observatory, Palisades, New York 10964*
THOMAS KEFFER* *Woods Hole Oceanographic Institution, Woods Hole, Massachusetts 02543*

ABSTRACT

A review of paleoceanographic studies dealing with late Quaternary deep-water circulation in the oceans is presented. These studies, which are based on the analysis of deep-sea benthic foraminiferal faunal and geochemical data and sedimentological data, are discussed in light of physical oceanographic conditions in the modern ocean.

We include in this review a re-evaluation of benthic foraminiferal data from North Atlantic and Southern Ocean piston cores and suggest that the dominance of *Uvigerina* during glacial intervals reflects increased amounts of organic carbon at the sea floor compared to modern values. A number of geochemical studies have suggested that the production or characteristics of North Atlantic Deep Water (NADW) changed during glacial times. Although these changes have been thought to result from a cessation of overflow water from the Norwegian-Greenland Seas, it is suggested here that seasonal sea-ice cover was possible over southern portions of the Norwegian Sea during glacial intervals. The presence of seasonally open water would have allowed Norwegian Sea Overflow Water to have been produced, although perhaps at lower volumes and with different hydrographic properties than at present.

The record of Antarctic Bottom Water (AABW) circulation does not show a simple relationship with paleoclimatic oscillations, indicating that changes in oceanographic conditions in the Southern Ocean had little effect on AABW formation. The AABW record contrasts with the glacial-interglacial cycles of NADW, suggesting no direct link between AABW and NADW circulation.

A variety of data suggests that changes in Pacific Deep Water circulation occurred as a result of glacial production of North Pacific deep water or from an increased flux of Southern Ocean water.

*Present address: School of Oceanography WB-10, University of Washington, Seattle, Washington 98195.

INTRODUCTION

High-latitude glaciations during the late Quaternary dramatically changed sea-surface conditions in polar seas relative to those found in the present day. As these regions are presently the formation sites of deep-water masses, the changes influenced deep- and bottom-water circulation throughout the oceans. During the past decade, numerous studies have provided information on past circulation conditions in the deep sea. The objectives of these studies were to describe deep-water circulation during the Quaternary, to compare these reconstructions with modern circulation patterns, and to relate the inferred changes to the paleoceanographic conditions of the source regions where deep water is presently formed.

The objective of this paper is to provide a broad overview of the current status of deep-water paleocirculation studies in light of physical oceanographic processes influencing modern deep-water circulation.[1] The approach is to review and synthesize studies which have provided information on Quaternary deep-water circulation. We begin with a brief outline of some of the paleoceanographic methods used, which include faunal, isotopic, and geochemical analyses of benthic foraminifera and sedimentological indices. Following this, the paper is divided into sections dealing with different regions of the ocean. For each region, a discussion of its modern physical oceanography relevant to deep-water circulation is presented, followed by a review of late Quaternary circulation studies for the area.

[1]Due to the large number of studies discussed in this review, figures are used to illustrate only (a) basic paleoceanographic methods, (b) relevant physical oceanography, and (c) paleoceanographic data reinterpreted in this paper.

PALEOCEANOGRAPHIC METHODS

Benthic Foraminifera

One of the first methods used for the determination of past circulation conditions was the analysis of deep-sea benthic foraminifera. Streeter (1973) and Schnitker (1974) were the first to suggest that the modern distribution of deep-sea benthic foraminifera in the North Atlantic correlates with the distribution of deep-water masses. These studies were followed by a number of regional distribution studies (see Douglas and Woodruff, 1981). Although some of these studies show high correlations between assemblages or species abundances and bottom-water hydrographic properties (for example, Lohmann, 1978a), they have not proven to be consistent from one region to another, indicating that the ecological controls of deep-sea benthic foraminifera are much more complex than initially thought. Water mass reconstructions are hence presently limited by a lack of understanding of the ecological controls of the foraminifera.

It is now clear, however, that some of the variation between foraminiferal data and environmental variables is due to microhabitat preferences of benthic foraminifera. An analysis of living (stained) benthic foraminifera from a box core taken on the United States continental margin in 3,000-m water depth revealed vertical stratification of four species within the upper 15 cm of sediment (Fig. 1a) with species-specific microhabitat preferences (Corliss, 1985). Three of the species (*Melonis barleeanum, Chilostomella oolina, Globobulimina affinis*) were suggested to have infaunal habitats: 73% of the living fauna were found below 2 cm. The distribution of these foraminifera may not be directly controlled by overlying bottom-water conditions, but by physio-chemical conditions within the sediments.

Other species, however, do show consistent correlations with environmental variables thus far and are useful in reconstructions. Two such species which have received a great deal of at-

Geological Society of America Bulletin, v. 97, p. 1106–1121, 10 figs., September 1986.

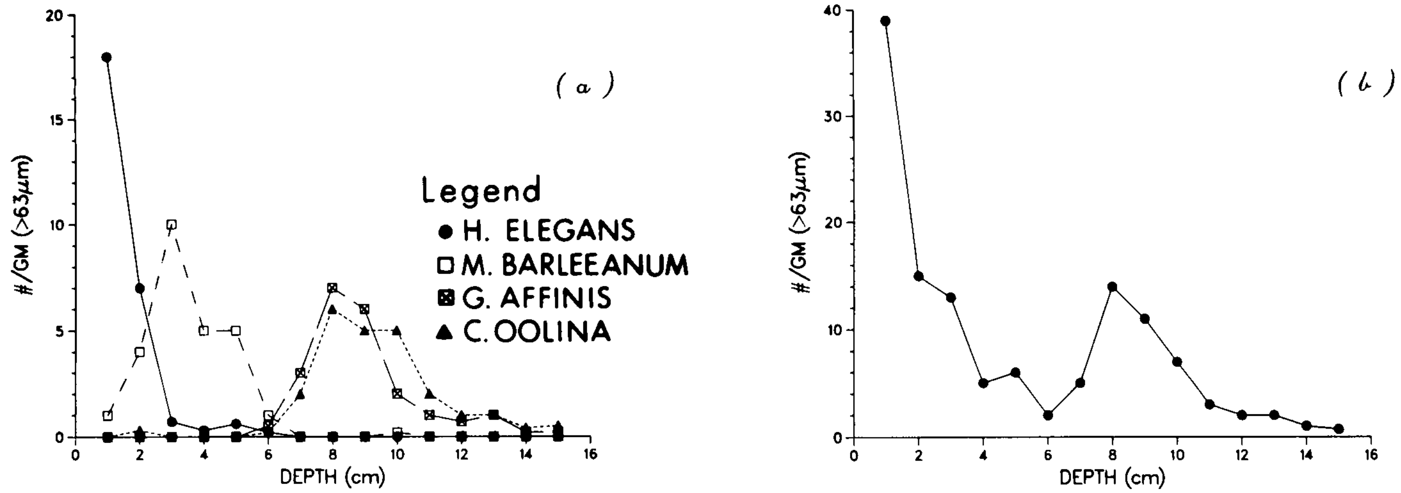


Figure 1a. The distribution of living (stained) benthic foraminifera within deep-sea sediments from box core Oceanus 86/2:7-4 taken from 3,000 m on the eastern United States continental rise. The data are expressed as number of foraminifera per gram of >63 μm sediment: (*a*) individual species data, (*b*) total number of foraminifera within each 100-cc sample (after Corliss, 1985).

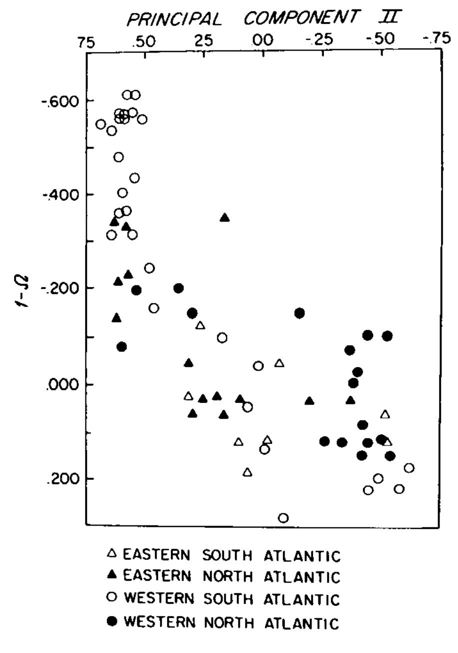
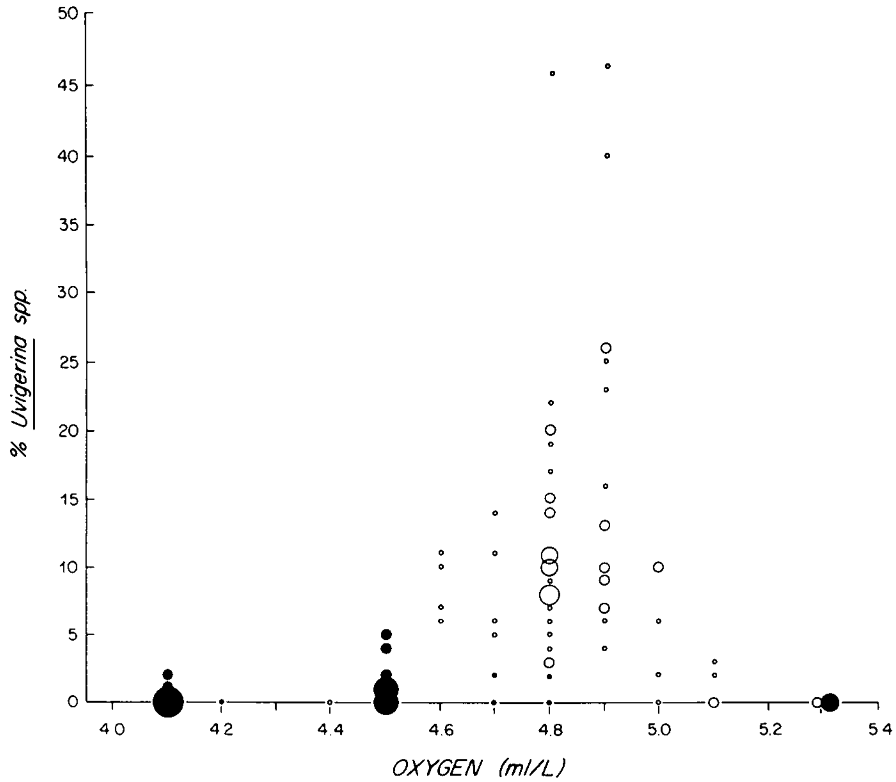


Figure 1b. The distribution of benthic foraminifera from Atlantic trigger core samples expressed as principal component II versus 1-Ω (Bremer and Lohmann, 1982). 1-Ω is an index of calcium carbonate undersaturation with the large positive values, indicating undersaturated bottom water. A positive loading of principal component II reflects a more diverse assemblage, whereas high-negative loadings reflect the dominance of *Epistominella umbonifera* alone. The data show that *E. umbonifera* is associated with highly undersaturated bottom water throughout the Atlantic Ocean.



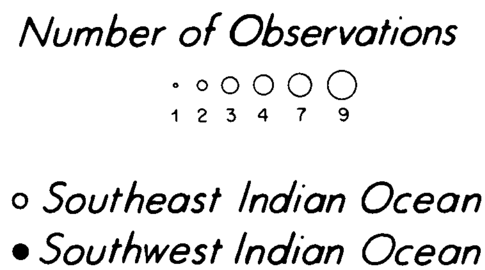


Figure 1c. The percentage of *Uvigerina* spp. versus bottom-water dissolved oxygen from the Southeast Indian Ocean (Corliss, 1979a) (open circles) and Southwest Indian Ocean (closed circles). The abundance data of *Uvigerina* in the Indian Ocean surface sediments do not correlate with dissolved oxygen content, as had been previously suggested (after Corliss, 1983a).

tention are *Epistominella umbonifera* and *Uvigerina peregrina*.

Epistominella umbonifera. A number of studies have suggested the importance of the undersaturation of calcium carbonate as an ecological influence on benthic foraminifera (Theyer, 1971; Murray, 1973; Greiner, 1974; Corliss, 1979a), but it was not until the work of Bremer and Lohmann (1982) that carbonate undersaturation and foraminiferal data were specifically compared. A correlation analysis of Atlantic Ocean faunal data and 1-Ω (an index of carbonate undersaturation), depth, and six hydrographic variables showed that the faunal data were best correlated with 1-Ω, depth, and bottom-water temperature. It was argued that the undersaturation of $CaCO_3$ (Fig. 1b) was the most reasonable explanation for the distribution of *E. umbonifera* in the deep sea. A similar pattern was observed in the Indian Ocean (Corliss, 1983a; Peterson, 1984). Based on existing data, the consistent association of *E. umbonifera* with highly undersaturated bottom water thus appears to provide a valid means to determine the record of undersaturation of $CaCO_3$ in the deep ocean.

Uvigerina peregrina. This species has been associated with old, low-oxygen bottom water throughout the ocean (Lohmann, 1978a; Corliss, 1979a; Burke, 1981; Streeter and Shackleton, 1979). Faunal and oxygen data from the Indian Ocean (Fig. 1c), however, reveals that the percentage of *Uvigerina* can vary independently of dissolved oxygen in the deep water (Corliss, 1983a). Instead, it appears to show a consistent relationship with high amounts of organic carbon and fine-grained sediment (Lutze, 1980; Douglas, 1981; Miller and Lohmann, 1982; Lutze and Coulbourn, 1984). For example, on the west African continental margin, high abundances (≥20%) of *Uvigerina peregrina* were associated with a wide range of bottom-water dissolved oxygen (2 to >5 ml/L), but only with relatively high organic carbon weight percentages of ≥2.0% (Lutze, 1980; Lutze and Coulbourn, 1984). On the North American margin, relative abundances of >80% were found with 0.75% to 1.0% organic carbon, and no relationship was shown with dissolved oxygen content of the bottom water (Miller and Lohmann, 1982). Further study is required to determine the microhabitat preference of *Uvigerina* to better understand the ecological requirements of this taxon.

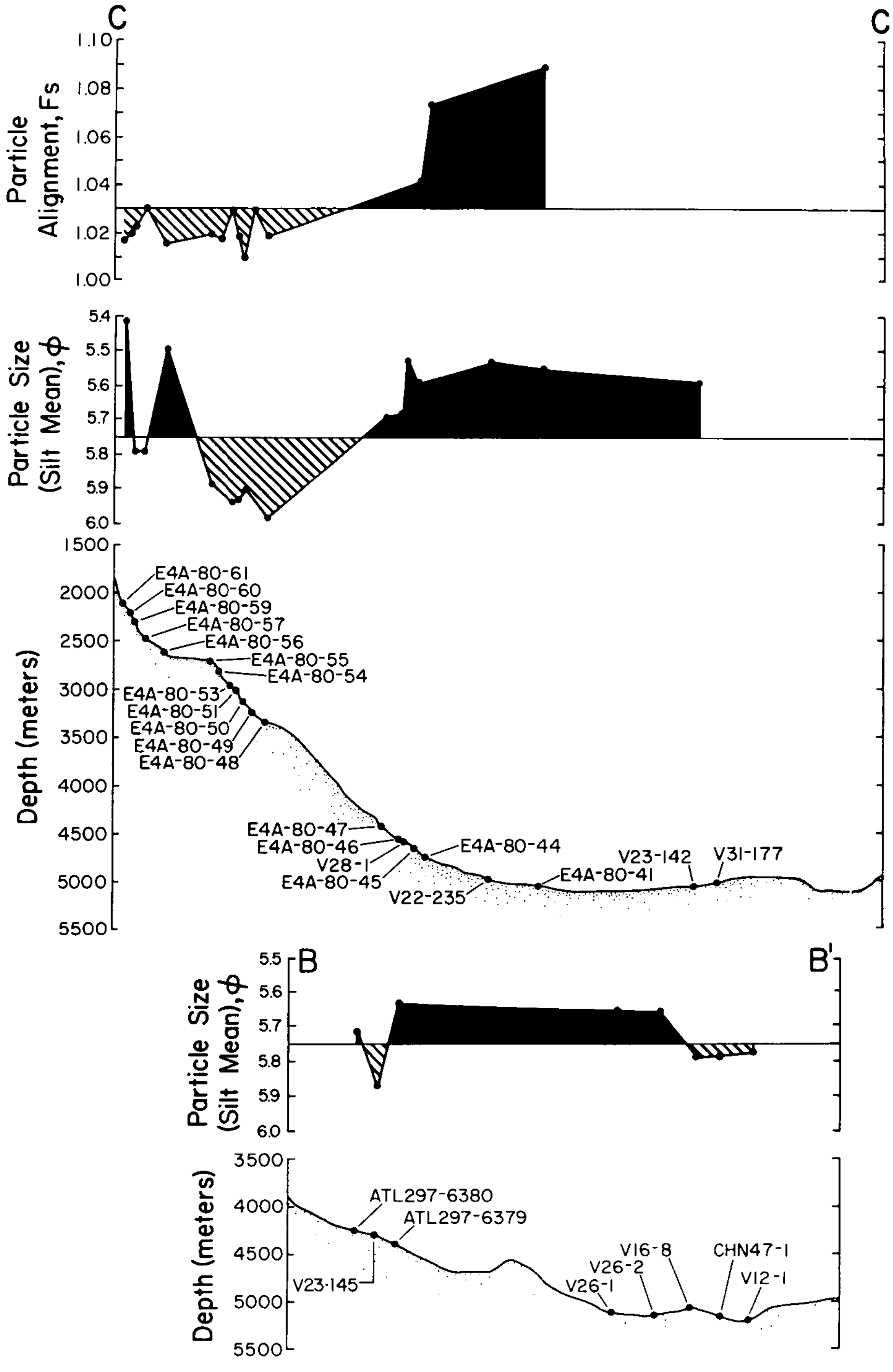


Figure 2. The mean noncarbonate silt size and particle alignment data from gravity core tops plotted against depth on three profiles of the United States east-coast continental margin between New Jersey and North Carolina. Increases in the particle size and particle alignment data are found between 4,400 and 5,200 m, which are suggested to reflect a high-velocity core of the Western Boundary Undercurrent (after Bulfinch and others, 1982).

Sediment Indices

The physical characteristics of deep-sea sediments have been used to evaluate relative bottom-water velocity changes in the deep sea. Grain size analysis of the noncarbonate silt fraction and fabric analysis (Fs parameter), determined by the measurement of anisotropy of magnetic susceptibility, were carried out on sediments from the Vema Channel beneath Antarctic Bottom Water (>4,200 m) and North Atlantic Deep Water (NADW) (<4,000 m) (Ledbetter and Johnson, 1976; Ellwood and Ledbetter, 1977; Ellwood and others, 1979; Ellwood, 1980; Ledbetter, 1984). Holocene sediment beneath the northward-flowing AABW was coarser and had higher Fs values (more uniformly aligned sediment grains) than sediment beneath the southward-flowing NADW, suggested to reflect a higher velocity flow of AABW than NADW.

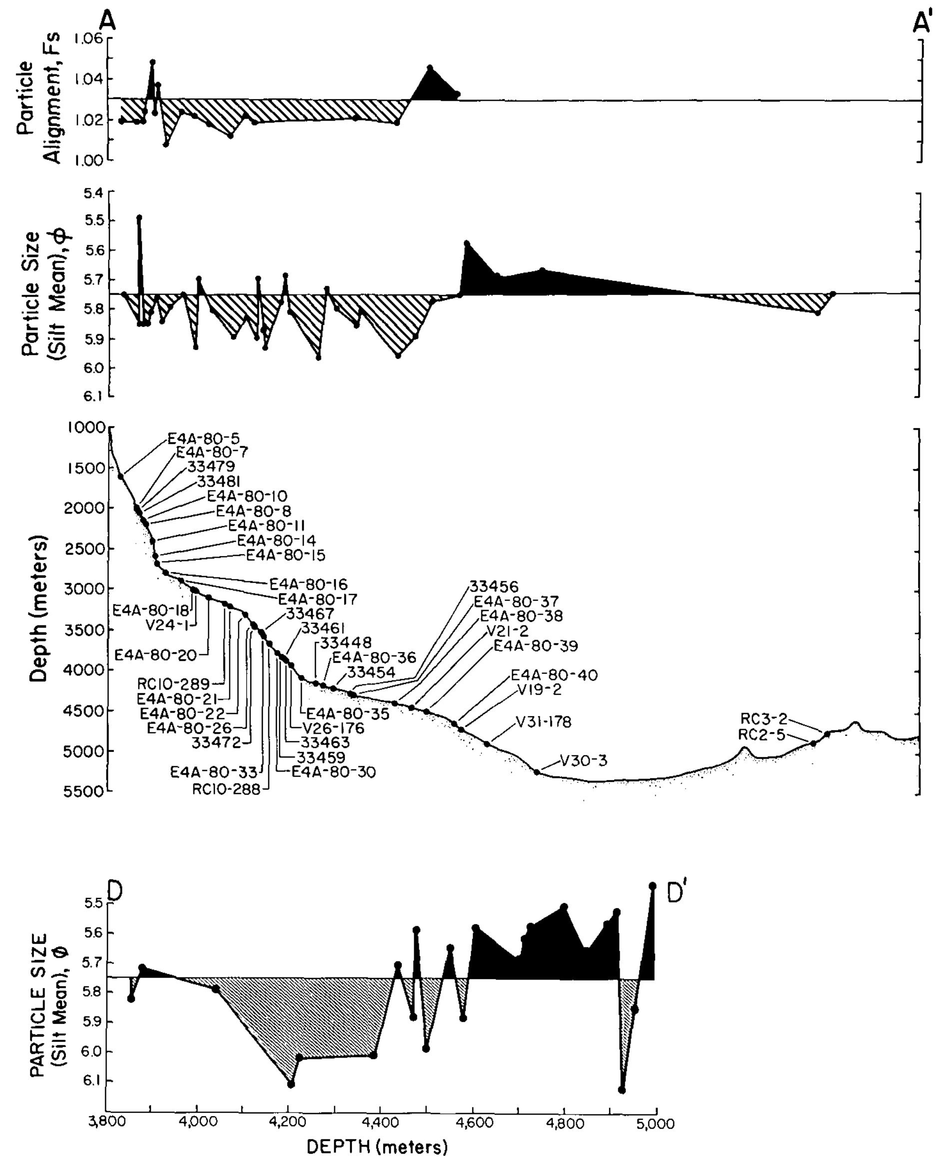


Figure 2. (*Continued*).

Additional studies were carried out on samples from the Amirante Passage (Indian Ocean) and the United States continental margin (Fig. 2). The sediment data reflected the presence of western boundary undercurrents (Ellwood and others, 1979; Ledbetter and Ellwood, 1980; Bulfinch and others, 1982).

The presence of displaced Antarctic diatoms in surface sediments has been used by Burckle and co-workers (Burckle and Biscaye, 1971; Burckle and others, 1973, 1974; Burckle and Stanton, 1975; Booth and Burckle, 1976; Johnson and others, 1977; Burckle, 1981) to determine the distribution of AABW in the Atlantic, Indian, and Pacific Oceans. Antarctic diatoms are entrained in AABW during its formation (Walsh, 1966), transported northward by AABW, and deposited along the path of bottom-water flow (Fig. 3). Analysis of diatoms in a depth transect of 23 core tops in the Vema Channel (Jones and Johnson, 1984) showed Antarctic diatoms to be abundant at depths >4,100 m, with the diatoms disappearing above 3,975–4,125 m, which corresponds to a level between the deeper northward-flowing AABW and southward-flowing NADW (Hogg and others, 1982).

Geochemical Indices

^{13}C Isotopic Data. Phytoplankton fractionate carbon isotopes during photosynthesis, resulting in organic matter with a relatively low ^{13}C content. The subsequent oxidation of this organic matter deeper in the water column increases the amount of ^{12}C over surface water levels (while decreasing dissolved oxygen content). This relationship can be seen in Figure 4a, with a comparison of AOU (apparent oxygen utilization, the difference between calculated saturation and *in situ* concentration of oxygen) and $\delta^{13}C$ of *Planulina wuellerstorfi* (Belanger and others, 1981; Graham and others, 1981). Hence, as a bottom-water mass moves farther from its source, its AOU will increase, and its $\delta^{13}C$ content will decrease. The ambient $^{13}C/^{12}C$ ratio is then recorded by benthic foraminifera which take up carbon to form their carbonate tests.

Cadmium/Calcium Data. Cadmium/calcium ratios of two taxa of benthic foraminifera, *Uvigerina* spp. and *Cibicidoides kullenbergi*, were compared by Hester and Boyle (1982) with estimated phosphate values from published deep-water data, and a close correspondence between Cd/Ca and phosphate was shown (Fig. 4b). It was suggested that Cd/Ca ratios of benthic foraminifera which incorporate cadmium into their tests could be used to estimate the nutrient content of deep water in the past, which in turn could be related to deep circulation patterns.

Oxygen Isotopes. Oxygen isotopic data of deep-sea benthic foraminifera have been used to infer bottom-water temperature fluctuations. The method is to compare an oxygen isotopic record from a particular core to a reference record from a location where it is assumed that the bottom temperature has not changed from glacial times. The glacial-interglacial ^{18}O amplitude of the reference core, which is assumed to be dominated by the ice-volume component, is subtracted from the glacial-interglacial amplitude of the core in question, and the difference is attributed to temperature variability (Duplessy and others, 1975). The assumption of constant bottom-water temperatures may or may not be correct, as we discuss below.

These methods have been used to determine past circulation conditions in the deep ocean. In the following sections, a summary of physical oceanographic processes relevant to deep-water circulation is presented for different regions of the ocean, followed by a review of paleoceanographic studies dealing with the area.

THE NORTH ATLANTIC

Physical Oceanography

The deep-water circulation is dominated by North Atlantic Deep Water (NADW) and Antarctic Bottom Water (AABW) in its southernmost regions. NADW is a complex mixture of four distinct end members with a broad range of temperatures and salinities (Fig. 5; Worthington, 1976): Norwegian Sea Overflow Water (NSOW), Labrador Sea Water (LSW), Mediterranean Sea Water (MSW), and AABW.

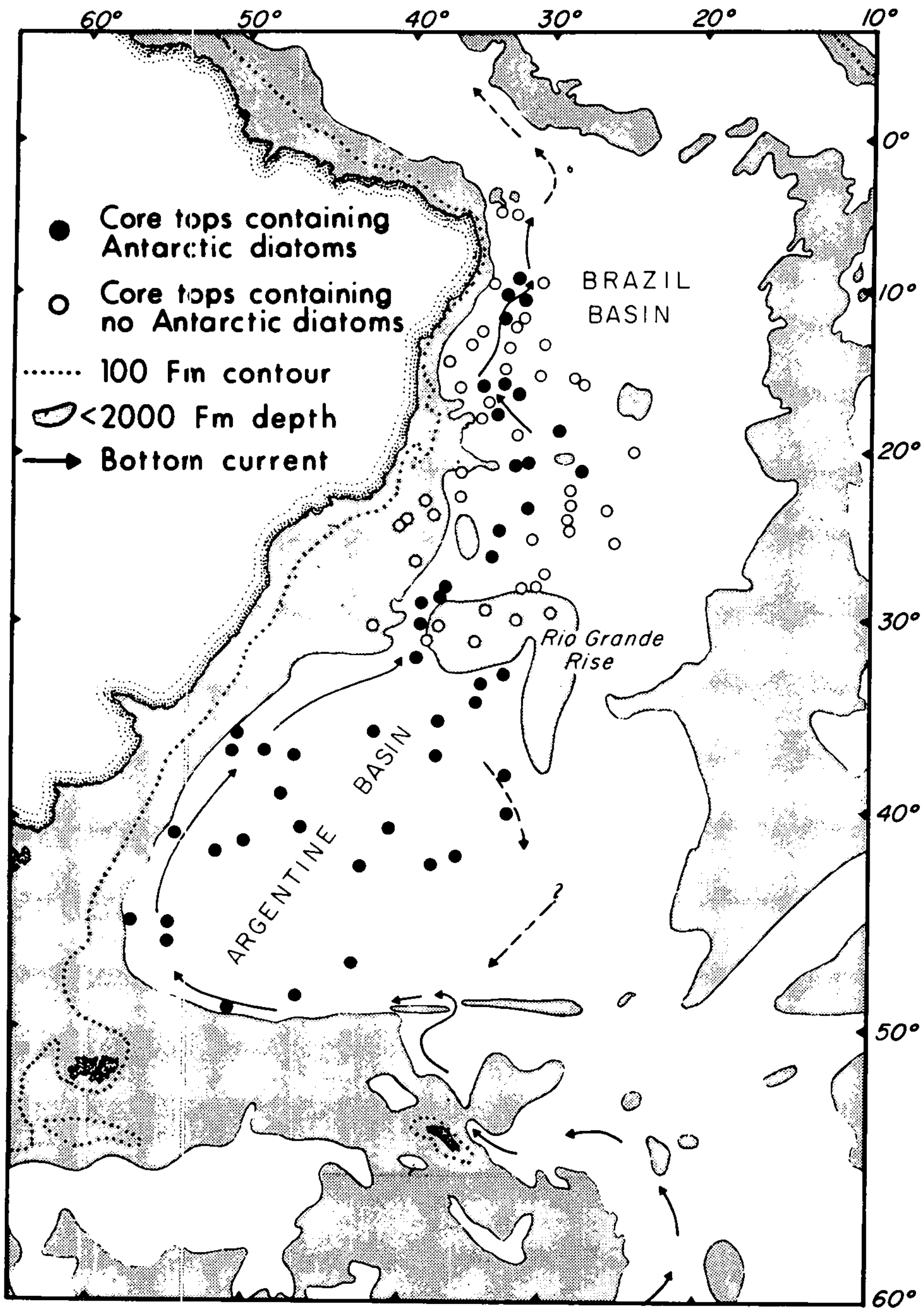


Figure 3. The distribution of displaced Antarctic diatoms in the southwest Atlantic Ocean. The shaded areas are shallower than 2,000 fathoms and the flow of Antarctic Bottom is indicated by arrows (Johnson and others, 1977).

Of the three NADW components originating in the North Atlantic, NSOW is the most dense due to its low temperature and relatively high salinity. The source of its saltiness has long been known to be from the south (Nansen, 1902), either from water advected into the Norwegian and Greenland Seas by the North Atlantic Current (Warren, 1983) or from Mediterranean Water drawn in by the overflow process itself or via isopycnal mixing along a density surface (Worthington, 1970; Reid, 1979; Saunders, 1982; Keffer, 1985).

Until recently, the deep waters of the Norwegian and Greenland Seas were thought to be the sources for NSOW. Worthington (1976) and Swift and others (1980), analyzing temperature and salinity data, and Peterson and Rooth (1976), analyzing GEOSECS tritium data, however, have concluded that it is the shallow water above the permanent pycnocline that is the principal source, not the deeper bottom waters. These surface waters are renewed by shallow open-ocean convection, making the present production mechanism of NSOW sensitive to atmospheric climatic fluctuations.

After being formed, NSOW exits the Norwegian and Greenland Seas through the Denmark Strait, over the Iceland-Faeroe Ridge, or through the Faeroe-Shetland Channel (Fig. 6). En route, the temperature, salinity, and transport of the overflow water is increased due to the entrainment of Atlantic lower thermocline water. In total, about 10 Sv (1 Sverdrup = 10^6 m^3/s) of overflow/entrained water exits the region to form NSOW. The NSOW enters the Labrador Sea to flow around the basin in a cyclonic sense at depths of 2,500–3,000 m. The NSOW exits the Labrador Sea and flows south to the Grand Banks, where its temperature (~2.2 °C), salinity (34.92‰), and density (σ = 37.098) closely resemble a cold and dense version of NADW (Fig. 5).

Labrador Sea Water, another important component of NADW, is formed as a result of open-ocean convection in the Labrador Sea (Clarke and Gascard, 1983), although here the convection is deep rather than shallow. The surface water that feeds this convection enters from around Cape Farewell (McCartney and Talley, 1982; Ivers, 1975). After it reaches the Labrador Sea, frigid winter air chills this water to about 3.3–3.4 °C, and continental runoff and an excess of precipitation over evaporation lowers its salinity to 34.84–34.89‰. Although open-ocean convection has never been observed in the Norwegian Sea, it was observed within the Labrador Sea by Clarke and Gascard (1983), who found a production of 105 km^3 of pure LSW during the winter of 1976. Their estimated average annual production rate of 3.9 Sv of pure LSW agrees well with an independent estimate of McCartney and Talley (1984), made on the basis of heat-budget calculations, of a total transport of 7.8 Sv, consisting of equal amounts of entrained and pure source water. This production is intermittent and may occur at 9–36 yr intervals (Talley and McCartney, 1982); it does not occur in years of low surface salinities or insufficient winter cooling.

After it is formed, the low salinity, highly oxygenated LSW resides at shallow to intermediate depths (to 2,500 m) and exits the Labrador Sea area either southward via the western boundary undercurrent, into the central North Atlantic where it mixes with Mediterranean Sea Water, or northward where it is entrained into the underlying NSOW (Talley and McCartney, 1982).

The Mediterranean Sea, the third North Atlantic source area for NADW, is unique because it is not located at high latitudes. The ba-

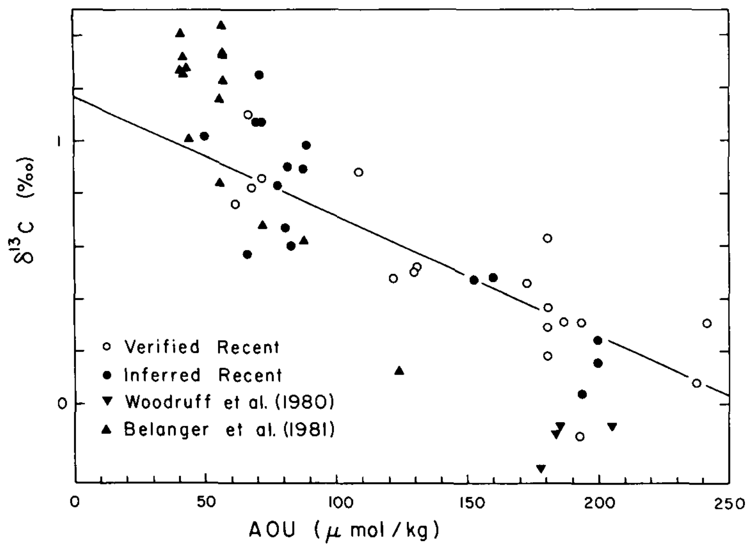


Figure 4a. The $\delta^{13}C$ composition of *Planulina wuellerstorfi* from core-top sediments from around the world ocean, plotted against apparent oxygen utilization (AOU, see text) from each core top location (Graham and others, 1981). An inverse relationship is seen, reflecting the dilution of ^{13}C due to the addition of ^{12}C to the bottom waters from oxidation of organic matter as the water mass flows from the North Atlantic into the Pacific Ocean.

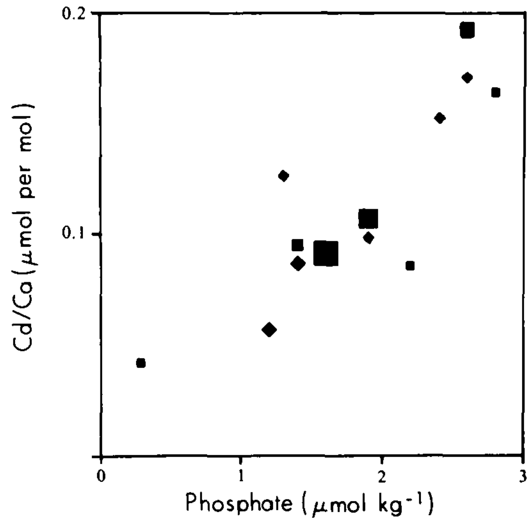


Figure 4b. The cadmium/calcium ratio of two species of Recent benthic foraminifera (■ *Uvigerina* spp., ◆ *Cibicidioides kullenbergi*) plotted against estimated bottom-water phosphate concentrations (after Hester and Boyle, 1982). The number of replicates is indicated by the size of the symbols. The data demonstrate a positive correlation between Cd/Ca content of benthic foraminifera and the phosphate content of the bottom waters.

sin's restricted inlet effectively traps the water, which undergoes substantial evaporation before it escapes. The resulting dense, salty water overflows the shallow sill at Gibraltar into the Atlantic, where it sinks to ~1,500-m depth, entraining substantial amounts of Atlantic water.

The degree of mixing of these NADW end members will influence the T-S characteristics at any particular location. For example, Jenkins and Rhines (1980), using tritium data, describe a ten-fold dilution of pure NSOW as it flows from the overflow region to the Blake-Bahama Outer Ridge (30°N). The amount of mixing between the various end members will be dependent upon distance from source locations, volume of water masses, and mixing mechanisms. Any relationship between observed properties in the basin to "upstream" properties at the source locations will be highly tenuous, because of the dominance of entrained (mixed) water away from the source regions.

NADW dominates the interior deep and bottom waters southward to about 15°N, where the final recognizable vestiges of AABW flowing northward are still sufficiently dense to form bottom water in the western basin. This AABW enters the North Atlantic as an eastern boundary current from the south along the western flank of the Mid-Atlantic Ridge (McCartney and others, in press; Warren, 1981a) and eventually mixes with the overlying NADW, losing its identity while further modifying the NADW.

The eastern basins of the North Atlantic are dominated primarily by NADW entering from the western basin through the Romanche Frac-

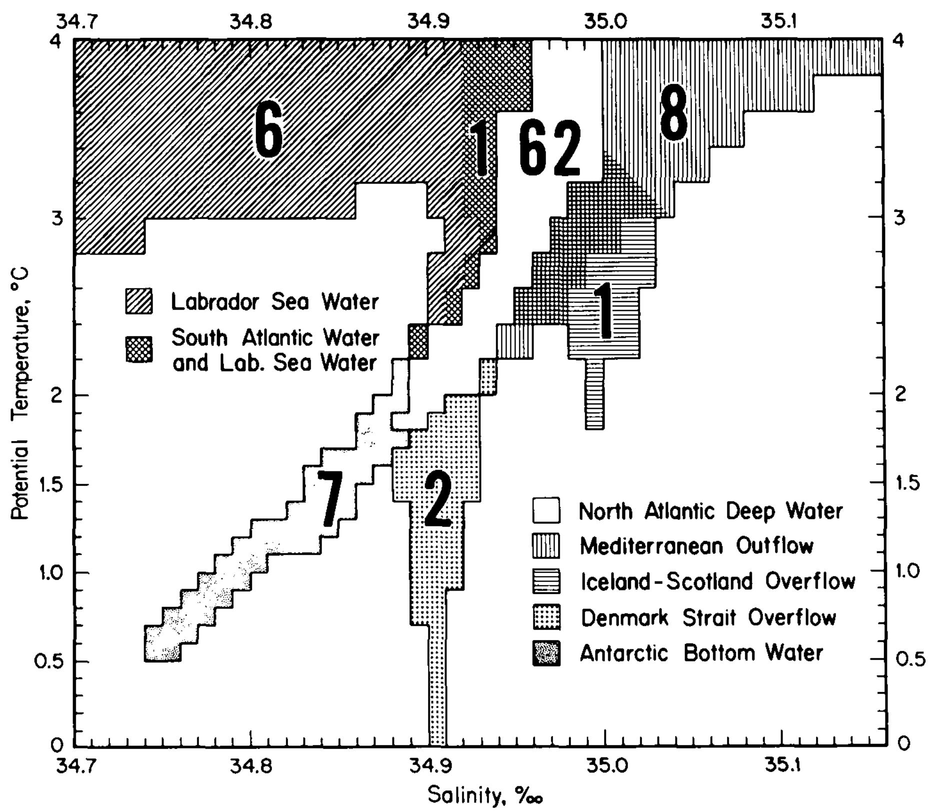


Figure 5. Temperature-salinity relationships of deep-water masses in the North Atlantic (from Worthington, 1976).

ture Zone and the Vema Fracture Zone, as well as NSOW entering directly from the Norwegian Sea. The sill depth of the Romanche Fracture Zone, the deepest fracture zone, lies at ~3,750 m (Metcalf and others, 1964). NADW flows into the eastern basin as a very nearly homogeneous water mass with potential temperatures of about 1.8–2.1 °C and salinities of 34.86–34.91‰ (Wright and Worthington, 1970). The upper limit of AABW in the western basin is below 3,750 m, and so little of it enters the eastern basin.

Bottom-water temperatures and silica concentrations increase to both the north and south away from the Romanche Fracture Zone (Warren, 1981a). The Sierra Leone Rise acts as a barrier to the north, preventing water < 1.8 °C (see Hobart and others, 1975) from filling the Gambia basin, and the Walvis Ridge acts as a barrier to the south.

The last component of bottom water in the northeastern Atlantic, NSOW, passes southeast of Iceland, either over the Iceland-Faeroe Ridge or through the Faeroe-Shetland Channel, to travel southward along the eastern flanks of the Mid-Atlantic Ridge as a relatively intense deep western boundary current. At the Charlie Gibbs Fracture Zone (53°N), it passes through the Mid-Atlantic Ridge and into the western basin. The influence of overflow waters at abyssal depths is generally confined to north of 50°N (see Worthington and Wright, 1970), although Broecker and others (1985) show this water as far south as 48°N.

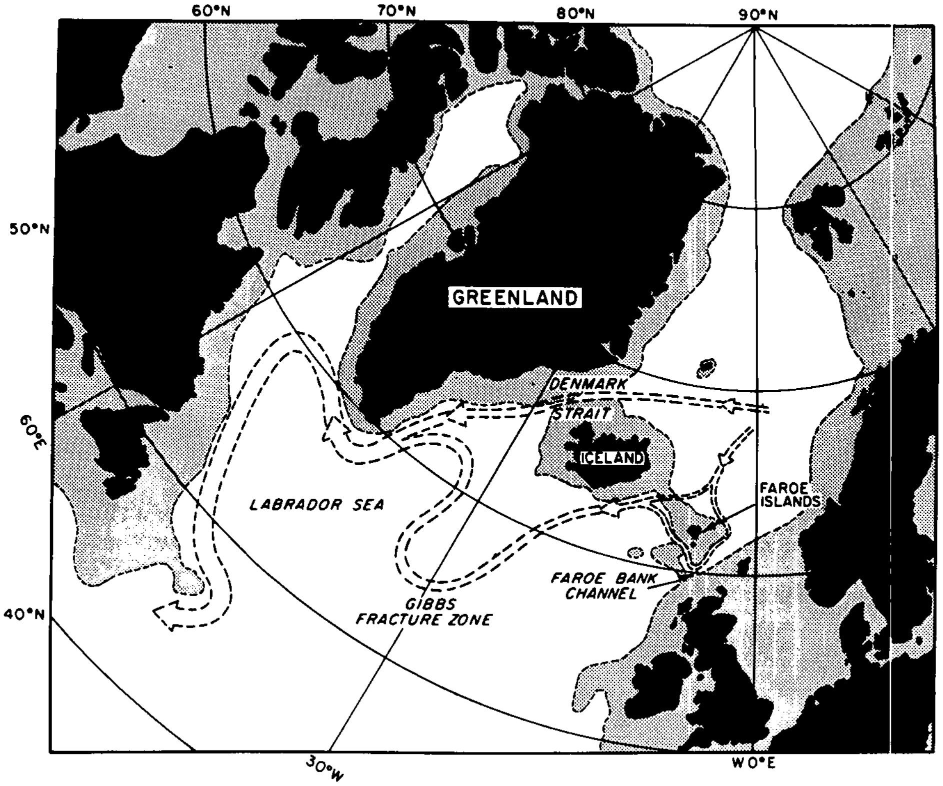


Figure 6. Deep-water paths for the Norwegian Sea Overflow Waters (from Warren, 1981a).

The Glacial North Atlantic

The Norwegian Sea. The role of the Norwegian-Greenland Sea in the production of overflow water during the late Quaternary has received widespread attention. In a series of papers, Kellogg (1975, 1976, 1977; Kellogg and others, 1978) suggested that during the past 450,000 yr permanent sea-ice cover was present most of the time over the area, and ice-free conditions existed only during the Holocene and in the last interglacial at about 120,000 yr B.P. It was suggested that an ice cover prevented deep-water mass formation, resulting in a cessation of saline NSOW that is presently contributing to NADW.

In a more recent paper, Kellogg (1980) suggested that the entire Norwegian Sea was covered by perennial sea-ice at 18,000 yr B.P., in contrast to conditions at 82,000 yr B.P., when seasonal ice covered the southern Norwegian Sea and permanent pack ice covered the remainder of the Norwegian and Greenland Seas. NADW was suggested to be formed during interglacial times and intermediate glacial conditions, such as those at 82,000 yr B.P., although in smaller amounts and on a seasonal basis, but ceased to be produced during full glacial periods. A similar conclusion was reached by Belanger (1982) who suggested ice-free conditions during the present interglacial, stage 5e, and at least partially ice-free conditions during stage 5a and stage 4.

Benthic foraminiferal oxygen isotopic data from piston cores were used by Duplessy and others (1975) and Streeter and others (1982) to make inferences about circulation conditions in the Norwegian Sea during the last glacial interval. Duplessy and others (1975) presented oxygen isotopic results from core K11 (water depth = 2,900 m) in the Norwegian Sea and compared the glacial-interglacial (stage 2 to stage 1) $\delta^{18}O$ amplitude with isotopic data from V28-238 (Shackleton and Opdyke, 1973) in the west equatorial Pacific. It was argued that the 1.1‰ greater amplitude between stages 1 and 2 in V28-238 could not be entirely accounted for by a cooling of Pacific Deep Water at 3,000 m overlying V28-238, because this would require Pacific deep waters to be 4.5 °C cooler during glacial stage 2 than the present-day 1 °C. Instead, it was suggested that Norwegian deep water was warmer than modern temperatures during part of stage 5 and all of stages 4, 3, and 2. This was hypothesized to be due to a cessation of deep-water formation at the surface of the Norwegian Sea and to an influx of North Atlantic water. Similar conclusions were presented by Streeter and others (1982) based on oxygen-isotope data from V27-86 (2,900 m).

These results, however, cannot be applied directly to Norwegian Sea Overflow Water which, as mentioned, originates *above* the pycnocline. Deep-water formation may have stopped, resulting in a warming of bottom waters and accounting for the isotopic results, but shallow convection and production of overflow water could have continued in the upper part of the water column, much as it does today. This overflow water, however, could not have been very dense, or it would have displaced the warm deep water. Low densities imply low salinities, which could have resulted from southern migration of the North Atlantic Current (CLIMAP, 1981), a cessation or reduction of Mediterranean Water into the Norwegian Sea, or seasonal addition of low-salinity meltwater.

Was the Norwegian Sea a Source of Glacial Deep Water? The Norwegian Sea is thought to have been covered by permanent sea-ice during 18,000 yr B.P., accompanied by a cessation of NSOW production. These conclusions were based on planktonic foraminiferal data and the percentage of calcium carbonate in surface sediments, with more emphasis put on the faunal composition data rather than on faunal abundance or carbonate content. Using the

Imbrie-Kipp (1971) method, the faunal composition data predict winter sea-surface temperatures of –0.5 to –2.1 °C and summer, resulting in a surface temperature of 2.6 to 4.9 °C at 18,000 yr B.P. These estimates are subject to considerable uncertainty and should be discounted, however, because the transfer function used to make them includes a monospecific factor, sinistrally coiled *Neogloboquadrina pachyderma*, which dominates Norwegian Sea sediments except during interglacials (Kellogg, 1980). Furthermore, the presence of *N. pachyderma* (sinistral) indicates polar sea-surface temperatures, but it is not directly related ecologically to the presence of sea-ice.

Areas covered with seasonal sea-ice in the present-day Norwegian Sea have calcium carbonate values between 10% and 30% and foraminiferal abundances between 1,000 and 10,000 specimens per gram, whereas regions with permanent sea-ice cover have calcium carbonate percentages of <10% and foraminiferal abundances of <1,000 specimens/gram (Kellogg, 1980). At 18,000 yr B.P. during full glacial conditions, calcium carbonate values >10% extended to as far as 78°N in the Norwegian Sea, and planktonic foraminiferal abundances ranged from 1,000 to 4,000 specimens per gram in the south-central Norwegian Sea (Kellogg, 1980). Both data sets suggest that seasonal (rather than permanent) sea-ice could have been present in sections of the Norwegian Sea.

The faunal and carbonate data at 18,000 yr B.P. fall within the lower range of values associated with seasonal sea-ice and are difficult to interpret because they may be influenced by a number of processes. For example, the modern values may be depressed due to dilution of terrigenous material, and bioturbation may have decreased modern carbonate values or enhanced 18,000 yr B.P. values. On the other hand, 18,000 yr B.P. carbonate data could have been depressed by increased carbonate dissolution and not by decreased surface productivity due to ice cover (Olausson, 1981).

With these reservations in mind, our interpretation of the 18,000 yr B.P. data is that sea-surface conditions were dramatically different from those of the present day, as suggested by Kellogg (1980), but seasonal ice cover was possible over southern portions of the Norwegian Sea during the entire glacial period. The presence of seasonally open water in the Norwegian Sea would allow NSOW to have been produced during the last glacial interval, although perhaps at lower volumes and with different hydrographic properties than at present. The presence of "warm" bottom water in the deep portion of the Norwegian Sea at 18,000 yr B.P. (Duplessy and others, 1975) would require that the paleo-NSOW was even less dense than the present surface-derived overflow water. A less-dense version of NSOW would have changed the over-all density of NADW. We suggest that the available evidence does not warrant the conclusion that production of overflow water from the Norwegian Sea ceased entirely during glacial intervals.[2]

The High-Latitude North Atlantic. Paleocirculation studies of the North Atlantic have attempted to determine whether or not deep circulation during glacial intervals was different from the modern circulation. As NSOW with present-day temperatures and salinity characteristics was most likely not being produced during full glacial times, a change in water-mass circulation would be expected to occur (Weyl, 1968). A variety of data shows that changes in deep-water advection and bottom-current patterns did indeed take place.

Oxygen. Oxygen isotopic data from piston core CH73139C in the northeast Atlantic were compared with data for the past 75,000 yr from southern Indian Ocean core MD73025 (Duplessy and others, 1980), and it was found that the glacial-interglacial amplitude in the North Atlantic core was 0.3‰ greater. It was suggested that the greater amplitude in the North Atlantic resulted from cooling of North Atlantic deep water by 1.3 °C during glacial times relative to the present due to deep-water convection in the North Atlantic. It was thus concluded that the North Atlantic was a source of deep water during the past 75,000 yr. The Circumpolar Deep Water (CDW) overlying MD73025, however, is composed of a mixture of Weddell Sea Water, NADW, and some water from the Indian and Pacific Oceans. An increase in temperature of any of these is possible, resulting in an increase of CDW temperatures which would, in turn, account for some of the 0.3‰ difference between these cores.

Cd/Ca. Cadmium/calcium ratios in one piston core (CH 82-31-11PC) from the western Atlantic covaried with oxygen-isotope data during the past 200,000 yr, with high Cd/Ca ratios during stages 2, 3, and 6 but low values during stages 1, 5, and the last part of stage 7 (Boyle and Keigwin, 1982). These data were interpreted to reflect changes in the nutrient content of the deep water on a glacial-interglacial cycle, reflecting a reduced flux during glacial intervals of nutrient-depleted NADW relative to that of Antarctic water. It was suggested that a cessation of NADW had not occurred. This conclusion was supported in a recent study of Cd/Ca and $\delta^{13}C$ of benthic foraminifera from a western North Atlantic core (CH 82-24-4PC) and an equatorial Pacific core (K 73-4-3PC), which suggested that a continuous net flux of nutrient-depleted water always occurred from the western North Atlantic to the Pacific during the past 215,000 yr (Boyle and Keigwin, 1985).

The late Quaternary Cd/Ca ratios may be influenced by regional surface productivity variations affecting nutrient loading (and, in turn, Cd/Ca ratios) on a glacial-interglacial cycle, rather than a pure physical oceanographic change. Furthermore, as Boyle and Keigwin (1982) pointed out, their conclusion on the flux of NADW is relative to the flux of southern water, and that this may not have been constant.

Carbon. Changes in $\delta^{13}C$ of deep waters through time could result from changes in terrestrial carbon flux to the ocean (Shackleton, 1977) or changes in the carbon/phosphorus ratio of sea water (Broecker, 1982). In the absence of any change in circulation, these global changes in $\delta^{13}C$ should be of the same magnitude everywhere and at all depths in the ocean (Curry and Lohmann, 1982). By comparing data from the same time intervals, but different water depths, Curry and Lohmann (1983) were able to disregard any global changes in the ^{13}C reservoir. Analysis of $\delta^{13}C$ of benthic foraminifera from the Sierra Leone Rise showed that glacial-interglacial differences were greater in cores below the sill depth (Romanche Fracture Zone) of the basin relative to those above the sill, implying a reduction in advection of deep water into the basin and an increase in residence time of deep water within the eastern Atlantic during the last glacial maximum.

Carbon and oxygen isotopic data from an eastern North Atlantic core (Meteor 123392) were compared with data from an eastern equatorial Pacific core (V19-30) by Shackleton and others (1983). The modern carbon isotopic difference of about 1‰ between deep waters of the North Atlantic and Pacific decreased during stages 2, 3, 4, and 6. It was suggested that the eastern North Atlantic had a colder, less-oxygenated water mass during glacials than at present, and the oxygen content (and age) of the glacial water in the equatorial Pacific and eastern Atlantic were similar. These conclusions conflict with the recent findings of Boyle and Keigwin (1985), who suggested that the nutrient

[2]Recent studies of southeastern Norwegian Sea piston cores (Jansen and others, 1983; Sejrup and others, 1984; Jansen and Bjorklund, 1985) show barren zones without planktonic foraminifera prior to 13,000 yr B.P., which are interpreted to reflect permanent sea-ice over the area during glacial times. These data do not invalidate our interpretation of 18,000 yr B.P. sea-surface conditions, because all of these cores lie east of 5°W, whereas the cores with foraminifera and carbonate data (Kellogg, 1980) suggesting seasonal sea-ice at 18,000 yr B.P. are to the west between 5°W and 10°W.

content (and age) of deep water in the Atlantic and Pacific were never equal. They attribute the isotopic differences between their study and that of Shackleton and others (1983) to the analysis of different benthic foraminiferal species and the continental margin location of Meteor 123392. Mix and Fairbanks (1985) had results similar to those of Shackleton and others (1983), but they reached different conclusions. They suggested that North Atlantic glacial deep waters were formed with high preformed nutrient contents, which resulted from formation beneath extensive sea-ice.

In a somewhat different approach, Duplessy and Shackleton (1985) presented $\delta^{13}C$ data of benthic foraminifera from 41 cores throughout the oceans in 3 time slices. During glacial stage 6 (135,000 yr B.P.) bottom-water circulation was apparently weaker than during interglacial times; apparently some deep water formed in the high-latitude North Atlantic close to permanent sea-ice, but most ocean basins were filled with water formed in the Southern Ocean. Within the deglaciation between isotopic stage 6 and 5e (~127,000 yr B.P.), a cessation of North Atlantic deep water production took place due to water column stratification caused by the southward displacement of sea-ice, with some deep water formed in the Southern Ocean. During stage 5d (~107,000 yr B.P.), a time of rapid continental ice growth, an enhanced flux of NSOW into the North Atlantic was suggested.

Sediments. The southward-flowing western boundary undercurrent that exists at about 4,900 m depth along the North American continental rise is an important component of NADW circulation and hence is an ideal location for studying fluctuations of NADW. Ledbetter and Balsam (1985) analyzed silt mean particle size and foraminiferal indices and showed that a boundary current was present throughout the past 25,000 yr, although the depth and velocity of the current has apparently varied considerably. No evidence was found for stagnation of the undercurrent during the last glacial interval. The circulation was as vigorous during glacial times as at present, but the depth and velocity of the undercurrent varied.

One problem with the use of grain size data for inferring past water mass velocities is the influence of changing provenance of the sediments over a glacial-interglacial cycle. For example, the initial grain-size distribution of sediment in any particular deep-sea location (before any winnowing has occurred by bottom currents) may vary from a glacial to interglacial period due to changing source area, changes in oceanographic conditions affecting sediment transport, or changes in sea level which in turn would influence the final grain-size distribution of the sediment. Variation in the initial grain-size distribution of sediments during a glacial-interglacial cycle would be particularly likely on or near continental margins, such as in the study of Ledbetter and Balsam (1985). A drop in sea level associated with build-up of Northern Hemisphere glaciers would allow fluvial deposits to be delivered directly into the deep sea, rather than being sequestered in estuaries or on the continental shelf, as happens during interglacial intervals.

Most paleoceanographic studies in the North Atlantic have considered past variations in NADW, but a study of AABW circulation during the past 550,000 yr was carried out by Bremer (1983), based on benthic foraminiferal data from four cores from the Cape Verde basin. Based on the association of *E. umbonifera* and AABW, she suggested an increase in volume of AABW, relative to the present, within the basin at 120,000 to 180,000 yr B.P., 280,000 yr B.P., 525,000 yr B.P., and the major increase at 375,000 to 425,000 yr B.P. As *E. umbonifera* is best related to the undersaturation of $CaCO_3$, the periods of high *E. umbonifera* should be interpreted as times of increased undersaturation of $CaCO_3$. Bremer (1983) compared her data with studies by Lohmann (1978b) and Corliss (1979b) and noted a general correspondence in the timing and magnitude of abundance changes of *E. umbonifera.* This coincidence in faunal patterns in the three regions reflects ocean-wide changes in the undersaturation of $CaCO_3$.

Labrador Sea. A third possible source region for deep water in glacial times is the Labrador Sea. This area was suggested as a source region by Fillon and Duplessy (1980), based on oxygen isotopic analyses of benthic and planktonic foraminifera from two cores, HU 75-41 (2,381 m) and HU 75-42 (2,403 m). The oxygen isotopic results of benthic foraminifera show a 1.8‰ change between stages 1 and 2, which is 0.15‰ greater than the 1.65‰ effect estimated for the maximum ice growth for stage 2 by Duplessy (1978). The 0.15‰ excess was attributed to a 0.5 °C cooling of Labrador Sea bottom water from modern temperatures due to glacial deep-water convection in the Labrador Sea. The analysis of mixed species of benthic foraminifera, however, may have contributed to this isotopic difference.

In evaluating the role of the Labrador Sea as a deep-water source area, one question which is critical and controversial is whether or not open-water conditions existed in the Labrador Sea during glacial times. The region had been previously considered to be covered with sea-ice throughout the glacial interval (CLIMAP, 1981), but benthic foraminiferal and pollen data from two cores from the Labrador Shelf (Vilks and Mudie, 1978) suggest that portions of the Labrador Sea may have been ice-free as long ago as 22,000 yr B.P., with regional presence of terrestrial vegetation. Fillon and Duplessy (1980) showed the presence of subpolar planktonic foraminifera in HU 75-41 and HU 75-42 during portions of stage 2, which they suggested reflected open-water conditions in a portion of the Labrador Sea during the glacial maximum. Mudie and Aksu (1984) suggested that the presence of foraminifera (>63 μm), dinoflagellates, and pollen in a Baffin Bay core at 66°N during portions of glacial stages 2, 4, 6, and 8 also indicated extensive areas of open water during the summer seasons.

These conclusions are contested on two points by Kellogg (1984, 1985). First, ^{14}C dates on organic carbon and shells from Labrador continental shelf cores show differences of >10,000 yr due to the addition of older organic carbon (Fillon and Harmes, 1982). Kellogg (1985) points out that the effect of contamination by older organic carbon could make the ^{14}C date used by Vilks and Mudie much too old. Secondly, the use of >63 μm foraminiferal data as an indicator of open-water conditions by Fillon and Duplessy (1980) was challenged (Kellogg, 1984), because >63 μm subpolar planktonic foraminifera were present at 18,000 yr B.P. in a core presently located beneath permanent sea-ice cover.

Changes in NADW circulation during glacial intervals have been suggested, yet the role of the *individual* sources of NADW is not identified. For example, V29-179, located east of the Mid-Atlantic Ridge, is overlain by bottom water with T-S characteristics typical of NADW. Early foraminiferal results from this core were interpreted to reflect changes in the production of Norwegian Sea Overflow Water (Streeter and Shackleton, 1979), and yet changes in any one of the four end members of NADW could have been equally responsible for the faunal changes. This illustrates the necessity of analyzing cores which are located at or near individual source regions, because cores taken "downstream" will be sensitive to changes in mixing dynamics, or in any one of a number of end members.

A Re-evaluation of Benthic Foraminiferal Data from the North Atlantic

Initial studies by Streeter (1973) and Schnitker (1974) noted that the distribution of benthic foraminifera changed in the North Atlantic during the late Quaternary. Subsequent studies have shown that the most distinctive feature during the last glacial interval is the wide-spread occurrence and abundance of *Uvigerina peregrina* (Schnitker, 1979; Streeter and Shackleton,

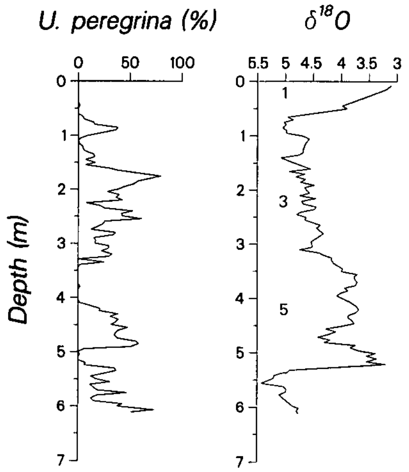


Figure 7a. Benthic foraminiferal isotopic data and the percentage of *Uvigerina peregrina* versus depth in V29-179 in the North Atlantic (after Streeter and Shackleton, 1979). The *Uvigerina* data show high abundances in stage 6, a portion of stage 5c, stage 5d, and stages 3 and 4. We interpret the high abundances of *Uvigerina* at these times to reflect increased amounts of organic matter at the sea floor relative to the present.

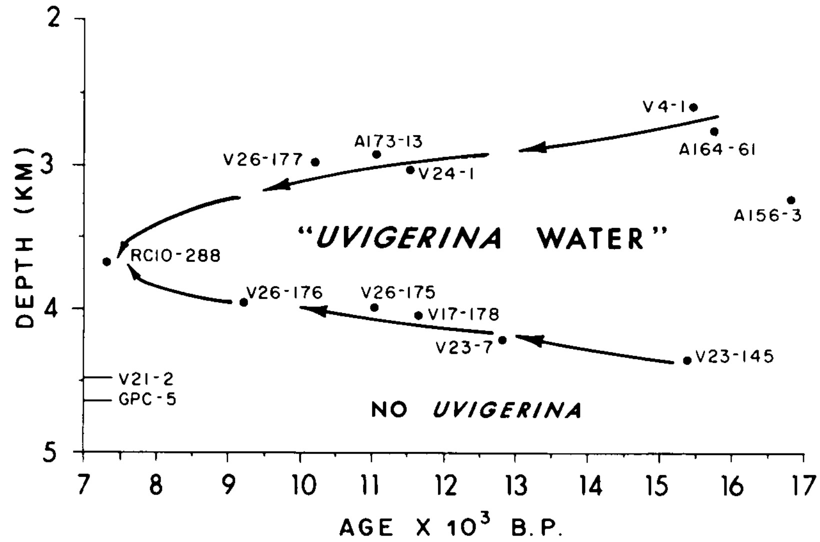


Figure 7b. The age of the last appearance of *Uvigerina* in sediment cores from the northwest Atlantic, plotted against the water depth of each core (after Balsam, 1981).

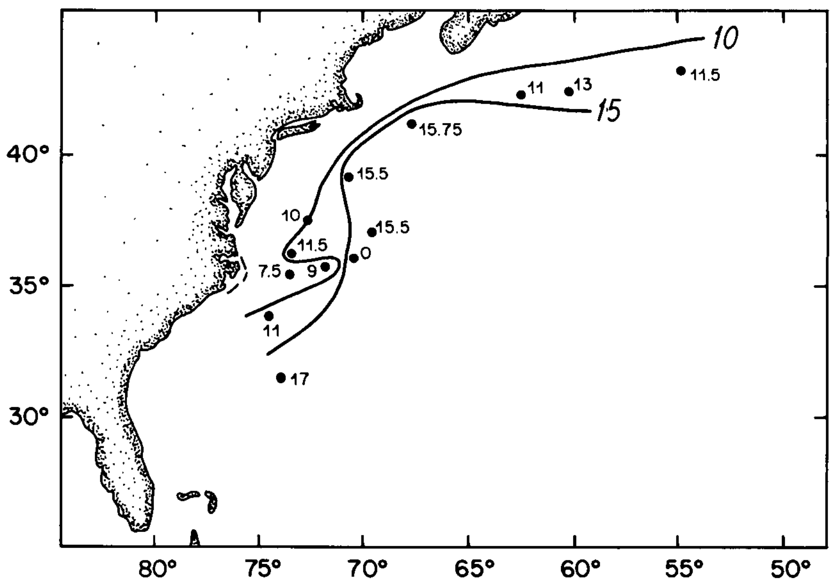


Figure 7c. The age of the last appearance of *Uvigerina* in sediment cores from the northwest Atlantic (Balsam, 1981) shown at the core locations. The last appearance data are contoured at 10,000- and 15,000-yr intervals and show a shoreward migration during the last deglaciation.

1979; Balsam, 1981; Streeter and Lavery, 1982). For example, Streeter and Shackleton (1979) analyzed piston core V29-179 (central North Atlantic) from oxygen-isotope stage 6 to 1 (150,000 yr B.P. to the present) and found a *Uvigerina-Melonis barleeanum-Astrononion* assemblage present during glacial stages 2, 3, 4, 5a, b, and in stage 6 (Fig. 7a). *Uvigerina* dominated during stage 6, a portion of stages 5c and 5d, and stages 3 and 4. This presence of abundant *Uvigerina* in glacial North Atlantic sediments was widely interpreted to reflect the presence of old, low-oxygen deep water due to a cessation or reduction of NSOW flowing into the North Atlantic. During the last deglaciation, *Uvigerina* disappeared in the western North Atlantic between 18,000 and 8500 yr B.P. with a mean age of 13,000 yr (Schnitker, 1979; Balsam, 1981; Streeter and Lavery, 1982). This diachronous disappearance was interpreted to have been caused by a replacement of old, glacial deep water with the renewed production of Norwegian Sea Overflow Water.

In evaluating the disappearance of *Uvigerina* in the North Atlantic, Balsam (1981) plotted the time of the disappearance in 14 northwest Atlantic piston cores against water depth of each core (Fig. 7b). He concluded that the disappearances did not have any geographical pattern but were related to water depth; he argued that this supported an interpretation of water-mass changes. In Figure 7c, however, the time of the *Uvigerina* disappearance at each core has been plotted at the core locations. Although the number of data points is small, a contouring of the data at 10,000 and 15,000 yr B.P. suggests that the disappearance of *Uvigerina* may indeed have a geographic pattern: disappearances occurred initially in cores farthest from land, migrating landward during the deglaciation.

On the basis of the correlation between abundant *Uvigerina* and relatively high organic carbon levels, we re-interpret the glacial *Uvigerina* data in the North Atlantic to reflect an increased amount of organic carbon at the sea floor which may have resulted from either increased surface-water productivity, increased river input of organic material directly into the deep sea due to the absence of continental shelves during lowered sea level, or enhanced preservation of organic matter due to low dissolved oxygen contents of the bottom waters (Emerson, 1985). The increase in organic carbon during glacial times would be similar to what was observed by Muller and others (1983) on the west African continental margin. It is not possible to determine which explanation best accounts for the inferred higher levels of organic carbon based on the existing faunal and sedimentological data. In any case, the *Uvigerina* data cannot be used at this time to argue directly for low oxygen conditions in bottom waters, because a first-order correlation between high abundances of *Uvigerina* and low oxygen levels has yet to be demonstrated.

THE SOUTH ATLANTIC

Physical Oceanography

The deep-water property distributions are dominated by southward-flowing NADW and northward-flowing Circumpolar Water (CPW) and AABW. Circumpolar Water "begins" in the South Atlantic, where NADW enters the Antarctic Circumpolar Current (ACC), and is carried eastward around Antarctica, all the while decreasing in O_2 and becoming fresher by mixing with Indian, Pacific, and Antarctic waters. As it re-enters the Atlantic through the Drake Passage, where it is near the bottom, it rises to 1,000 m (at 50°S, just below the Antarctic Intermediate Water) on the same isopycnal occupied by the southward-flowing NADW. This CPW spreads northward above and below the NADW (see Reid and others, 1977; Fig. 6); both upper and lower branches of the CPW are characterized by low O_2 which monotonically decreases to the north. The upper CPW lies at a 1,500-m depth in the subtropical gyre, and the lower CPW (represented by a O_2 minimum) lies at 3,500 m and is not seen north of the Rio Grande Rise.

The formation of AABW around Antarctica is a complex subject that has historically received considerable attention on mechanisms ranging from air-sea interactions to the effect of nonlinearities of the equation of state (cabelling). Traditionally, the continental shelves surrounding Antarctica have been accepted as the formation sites of the precursors of AABW (Brennecke, 1921; Mosby, 1934), particularly beneath the Filchner and Ronne Ice Shelves of the Weddell Sea.[3]

In the traditional model, production of sea-ice on the broad continental shelf of the southwest Weddell Sea produces a very cold ($T < -1.8$ °C) and salty ($>34.6‰$) shelf water which moves offshore and then mixes with a modified form of warm Weddell Deep Water (Foster and Carmack, 1976). The resulting water is then moved down from the shelf and onto the floor of the Weddell Sea (Gill, 1973; Killworth, 1977). What emerges is a water mass with T-S properties very similar to classical AABW ($T = -0.3$; $S = 34.66‰$). This mechanism, however, will work only when circulation on and off the shelf is possible. Should glacial ice be grounded on the shelf, this area will be unavailable for bottom-water production (Kellogg, 1986).

Recently, the discovery of the Weddell polynya (Zwally and others, 1976), combined with observations of a Weddell open-ocean convective "chimney" (Gordon, 1978), have led to another theory on the formation of AABW. NADW represents an input of warm salty water to the Southern Ocean south of the ACC. This water rises to shallow depths in the Southern Ocean due to the steeply sloping isopycnals associated with the ACC and the cyclonic Weddell gyre. The NADW influence thus is seen in the waters immediately below the pycnocline within the Weddell gyre. Although NADW is warm (~2 °C), its salt makes it more dense than the fresher surface waters of the Weddell gyre. Its density is only marginally greater than the surface waters, and so salt rejection during sea-ice formation can result in deep convection. This convection, which brings up the warm deep water to the surface where it melts the ice, is the mechanism thought to account for the occurrence of the Weddell polynya (Gordon, 1978; Martinson and others, 1981).

Modeling results of Martinson and others (1981) have shown that the modified water mass resulting from this convection is of the proper characteristics, volume, and depth to be a major component of AABW. A review of the historical data from the Weddell area by Gordon (1982) has demonstrated the existence of the water type predicted by the model. This led Gordon to speculate that during years in which open-ocean convection occurs (as indicated by the presence of the polynya), the resulting deep water may "short circuit" the traditional shelf mechanism of bottom-water production and produce the required deep-water mass directly. If this is true, bottom-water formation will be more dependent on local dynamics and atmospheric forcing than on local geography.

Regardless of its method of formation, the AABW must escape the Weddell Sea region and cross the ACC. Georgi (1981) suggested that a dense abyssal component, whose source is mainly Weddell Sea Bottom Water, flows northward through the South Sandwich Trench and Falkland Channel (Fig. 8). A slightly less dense, overlying component enters the Argentine basin. This is the more traditional AABW which Reid and others (1977) attribute to Weddell Sea Deep Water, which spreads northward via isopycnal mixing across the ACC. AABW is dominated by a western boundary flow which passes northward through the Vema Channel into the Brazil basin and eventually into the North Atlantic (Hogg and others, 1982; Whitehead and Worthington, 1982).

The water masses of the eastern South Atlantic, below the depth of the Mid-Atlantic Ridge, are essentially dominated by high-oxygen, low-temperature, and low-salinity water originating directly from the Antarctic. No strong deep circulation is evident in the eastern basin.

South Atlantic Paleoceanography

A number of studies have used Vema Channel cores taken on depth transects beneath NADW, Circumpolar Water, and AABW to determine paleocirculation conditions (Johnson, 1984 and reference therein). This area was chosen for this extensive paleoceanographic effort because it serves as a conduit for northward-flowing AABW.

Carbon isotopic analyses of the benthic foraminifer *Planulina wuellerstorfi* were carried out by Curry and Lohmann (1982) from seven piston cores to determine glacial-interglacial differences of ^{13}C. The shallowest and deepest cores showed no glacial-interglacial changes in $\delta^{13}C$, but four of the five cores located within NADW had $\delta^{13}C$ lower (implying a higher AOU) during glacial stages than at present. These lower $\delta^{13}C$ values were concluded to reflect a reduction in production of NADW during glacial intervals, because the lower $\delta^{13}C$ values were confined to the cores within NADW.

Mean particle size data of the carbonate-free silt fraction in three cores showed that the grain size data do not show a simple relationship with paleoclimatic conditions (Ledbetter, 1979, 1984). High relative AABW velocities were inferred to have occurred in stages 2 and 3, and at the stage 5/6 and stage 5/4 boundaries, with low velocities in stages 1, 5, and 6. The record from one of these cores, CH 115-61, was extended back to stage 12; it showed low relative velocities from stage 6 to stage 12 (Ledbetter and Ellwood, 1982). The particle alignment parameter (Fs) was generally higher with larger particle sizes from stages 1–5, but before stage 5, the Fs data do not follow the particle size data.

This work was extended to include three time slices (18,000, 120,000, and 140,000 yr B.P.) on the basis of data from 27 cores (Ledbetter, 1984). The results from the three time slices at all depths show that a simple relationship between deep water-mass velocities and climate simply does not exist (Ledbetter, 1979, 1984).

[3]As pointed out by Jacobs and Georgi (1979), however, through the years the number of "accepted" source regions has increased with the number of observations along the shelves.

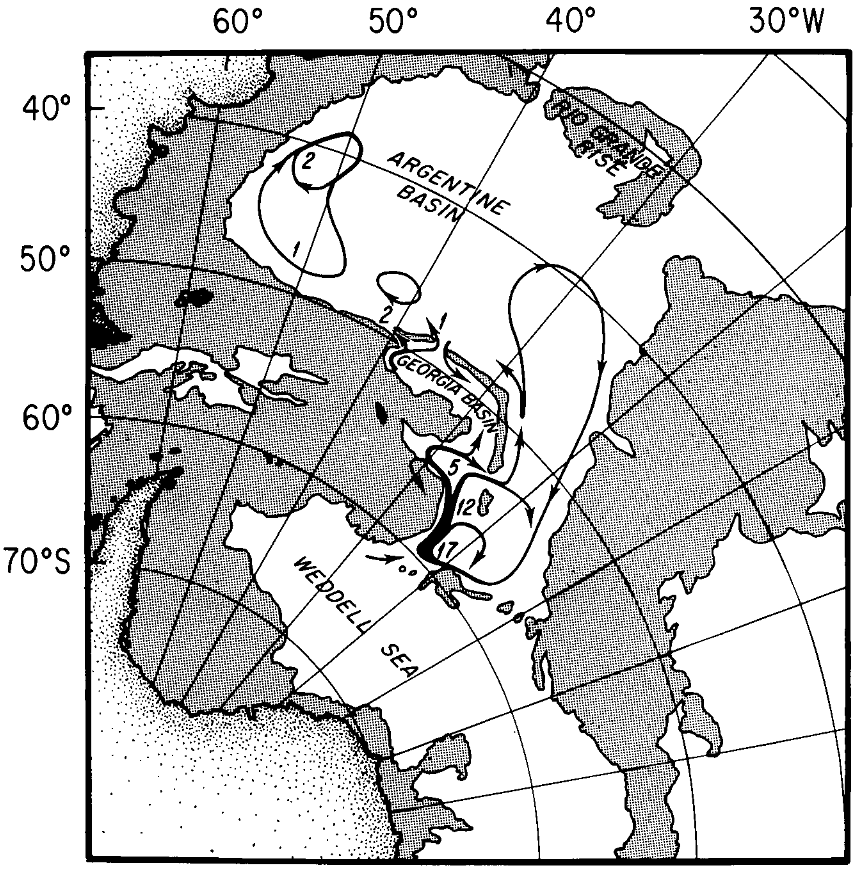


Figure 8. Schematic abyssal circulation for the southwestern South Atlantic. Numbers indicate baroclinic transports in units of 10^6 m^3 s^{-1}, relative to the 0 °C potential temperature surface, except in the eastern Argentine basin, where a deeper reference level is assumed. Shaded areas are less than 4,000 m deep (from Georgi, 1981).

Depth profiles of calcium carbonate content and planktonic foraminiferal fragmentation were related to bottom-water hydrography (Tappa and Thunell, 1984). Constant values from 1,500 to 3,900 m are found, but carbonate values decrease and the percentage of fragmentation increases in deeper samples. The 3,900-m level corresponds to the AABW/NADW transition; the sediment patterns could be related to the relative position of the water masses. Data from glacial stages 2 and 6 showed similar depth distribution to the present day, suggesting no significant change in water-mass positions during the past 140,000 yr.

Microfossils were the focus of two studies in the Vema Channel. An analysis of diatoms in piston cores (Jones and Johnson, 1984) showed peaks of displaced diatoms at isotopic stage boundaries 7/6 and 3/2. These peaks were suggested to reflect increased transport of AABW, although temporal variation in diatom productivity in the Antarctic region is another explanation that cannot be ruled out. Benthic foraminifera were studied by Lohmann (1978b), who found the relative abundance of *E. umbonifera* to fluctuate during the late Quaternary. From these data, an increase in carbonate undersaturation can be suggested to have occurred from 40,000–80,000 yr, 100,000–210,000 yr, and 350,000–480,000 yr B.P.

THE PACIFIC OCEAN

Physical Oceanography

A variety of cold (–0.5 °C), high-salinity (34.75‰) AABW is produced beneath the Ross Ice Shelf (Jacobs and others, 1970, 1979; Gordon, 1972a) which flows into the southeast Indian and southwest Pacific Oceans (Gordon, 1972a, 1972b; Rodman and Gordon, 1982). AABW enters the Pacific below the CPW and is most pronounced in the southwest western boundary current.[4] The northward-flowing boundary current can be followed as a salinity maximum until it passes through the Samoan Passage and into the North Pacific (Warren, 1981a), where this water apparently flows to the east (Mantyla, 1975) and northwest (Warren, 1981a). Although some evidence of a southward-flowing recirculation against Japan exists (Warren, 1981a), the deep water of the North Pacific is thought to circulate slowly around the large basin and back through the South Pacific, rejoining Antarctic Circumpolar waters.

The modern North Pacific Ocean has no source areas for deep or bottom water due to the lack of high-salinity surface water in high latitudes (Warren, 1983). Pacific deep water does show a distinct low-salinity signature which cannot be attained by simple mixing of North Atlantic and Antarctic source waters but is acquired through vertical mixing with the overlying North Pacific low-salinity waters. This mechanism of vertical mixing creates a unique deep water whose low salinity makes it a recognizable water mass (Gordon, 1983).

The Glacial Pacific Ocean

Pleistocene changes in the Pacific deep circulation are largely unknown. It is tacitly assumed that Pacific deep water remained unchanged from the present day, although some data suggest that this assumption is incorrect. Mangini and others (1982) reviewed ^{230}Th and ^{231}Pa data from North Pacific sediments and suggested that sediment erosion occurred over large areas at 70,000 yr B.P., near the oxygen-isotope stage 5/4 boundary, due to increased bottom-water flow. Seismic-reflection profiles from the northwest Pacific basin revealed unconformable, migrating or truncated sub-bottoms (Damuth and others, 1983) as well as sediment drifts and an erosional channel (Mammerickx, 1985), all thought to be formed by intense deep-water thermohaline circulation. Modern circulation of Pacific Deep Water in the area is sluggish, and so it was reasoned that strong bottom currents created these features in the past.

Carbon isotopic data from North and equatorial Pacific cores led Shackleton and Duplessy (1985) to suggest that the North Pacific was a source of deep water at 18,000 yr B.P. The North Pacific cores showed more positive $\delta^{13}C$ values than the equatorial cores, which indicate that the northern cores were overlain by younger deep water than that found overlying the equatorial cores.

There is little doubt that sea-surface conditions in the North Pacific were markedly different in glacial times (Fig. 9). These changes included the migration of the Subarctic Convergence (Polar Front) southward about 5° latitude

[4]B. A. Warren (1985, personal commun.) pointed out that it is difficult to consistently distinguish AABW and NADW in the Indian and Pacific Oceans. We use the term "AABW" here and in the Indian Ocean discussion to refer to water composed of some mixture of these two water masses.

(42° to 35°–38°N), a slight southward shift and increase in the surface-temperature gradient, a 6–8 °C cooling of surface waters in temperate regions, and a major increase in extent of sea-ice over the Bering Sea and the northwest Pacific (CLIMAP, 1981; Thompson and Shackleton, 1980; Thompson, 1981; Sancetta, 1979; Moore and others, 1980).

On the basis of reconstructions of sea-surface temperature and sea-ice distributions, a likely region for Pacific glacial deep-water convection would have been in the northwest Pacific between the Polar Front and the glacial sea-ice margin, to the north of the warm, saline Kuroshio Current. The surface waters in this region were thought to be 6–8 °C cooler at 18,000 yr B.P. than at present, and an increase in surface salinity could have resulted from a number of factors, including injection of salt due to sea-ice formation, reduced fresh-water run-off, the positioning of the warm, saline Kuroshio Current near a region with intense winter cooling, and high rates of evaporation.

A second possible explanation for changes in Pacific deep-water circulation is that the flux of Southern Ocean deep or bottom water increased during glacial times. This latter region had dramatically different sea-surface conditions during glacial times (CLIMAP, 1981).

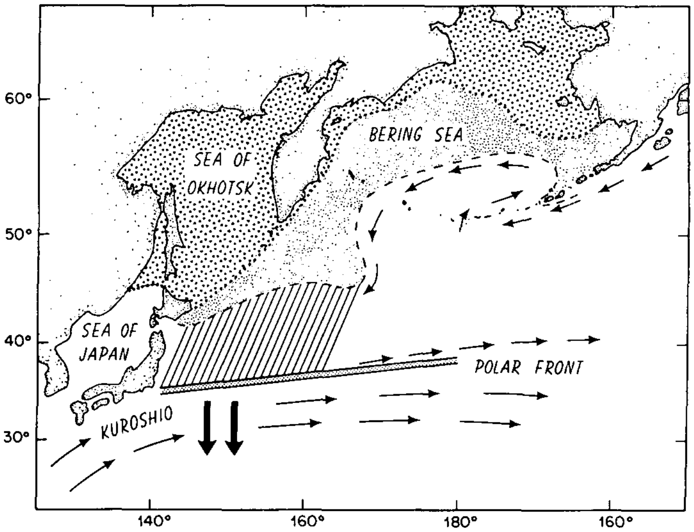


Figure 9. Schematic drawing of the northwest Pacific at 18,000 yr B.P. The locations of present-day (. . . . .) and 18,000 B.P. (- - -) winter sea-ice limit (CLIMAP, 1981) and locations of the Polar Front and the Kuroshio Current at that time (Thompson and Shackleton, 1980; Thompson, 1981) are also shown. Arrows in the Bering Sea indicate probable surface currents, and large arrows indicate possible deep-water flow from the northwest Pacific.

INDIAN OCEAN

Physical Oceanography

A network of major ridges divides this ocean into numerous basins. The most southwestern, the Crozet basin, contains AABW from the Weddell Sea which enters between the Crozet and Kerguelen Plateaus. This water passes through fracture zones in the Southwest Indian Ridge into the Madagascar and Mascarene basins where it forms a thin (500–700-km-wide) current below 3–4 km. It can be traced as far north as the Somali basin (Warren and others, 1966) via a distinct western boundary current with a structure quite similar to its southwest Pacific counterpart. Overlying and adjacent to this, there is a southward-flowing countercurrent of high salinity and silica but low-oxygen water originating from the northern Indian Ocean.

The Kerguelen Plateau and the Ninetyeast Ridge separate the Wharton, South Australian, and South Indian basins from the western basins, and prevent the flow into these regions of AABW originating from the Weddell Sea. AABW found in the eastern basins is believed to be a blend of bottom water produced near the Adelie Coast and from the Ross Sea (Gordon, 1974; Gordon and Tchernia, 1972). AABW mixes with Circumpolar Water as it travels from the Antarctic continental shelf through the South Indian basin, and it flows through the Australian-Antarctic discordance located at 120°–128°E in the Southeast Indian Ridge (Rodman and Gordon, 1982). The AABW flows into the South Australian basin and northward to the Wharton basin where Warren (1977) observed the flow as a narrow (600-km-wide) boundary undercurrent at a depth of 3,000–4,000 m.

The Central Indian basin is closed to the Southern Ocean below 4 km (Kanaev and others, 1977). Waters with AABW characteristics are observed in the eastern part of the basin, which originate as overflow water from the Wharton basin across several passages in the Ninetyeast Ridge. The AABW is then transported south with the general interior flow of the deep basin (Warren, 1981b).

Paleoceanographic Studies

Analysis of benthic foraminifera in one core, E48-03, taken beneath a western boundary undercurrent containing northward-flowing AABW in the South Australian basin (Corliss, 1979b), can be interpreted as reflecting changes in the undersaturation of calcium carbonate during the past 500,000 yr. The maximum corrosivity occurred during the equivalent of stage 11 (~420,000 yr B.P.); low corrosivity occurred primarily during the equivalent of stages 8 and 7 (305,000 to 195,000 yr B.P.).

Late Quaternary piston cores beneath the Antarctic Circumpolar Current contained a benthic foraminiferal assemblage dominated by *Uvigerina peregrina, Melonis barleeanum*, and *Melonis pompilioides* (Fig. 10; Corliss, 1982, 1983b). This assemblage was thought to reflect modification of Circumpolar Water during glacial intervals due to either a reduction or cessation of NADW circulation into the Antarctic Circumpolar Current. In light of the discussion of *Uvigerina* data in the North Atlantic, however, an alternative explanation is that these data reflect higher organic carbon levels at the core sites, resulting from higher surface-water productivity or enhanced preservation of organic matter.

THE GLACIAL SOUTHERN OCEAN

Micropaleontological and sedimentological studies suggest that AABW circulation is similar during glacial and interglacial intervals, although the data are not entirely consistent. The sedimentological data show no simple relationship between AABW circulation patterns and climatic fluctuations, whereas the displaced diatom

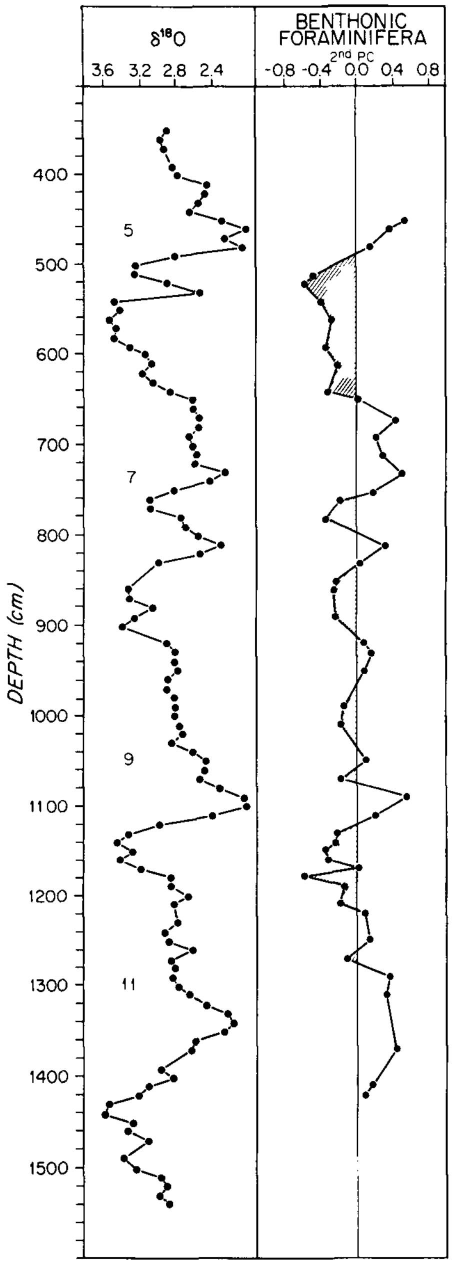


Figure 10. The second principal component of benthic foraminiferal data and planktonic foraminiferal oxygen isotopic stratigraphy (Hays and others, 1976) from E49-18 in the Southeast Indian Ocean sector of the Southern Ocean (after Corliss, 1982). Two benthic foraminiferal faunal assemblages are identified. One assemblage (negative loadings) indicates the dominance of *Uvigerina peregrina* (among others) and is found during glacial periods and intermediate cool intervals. A second assemblage (positive loadings) is found generally during warm interglacial intervals and is dominated by *Globocassidulina subglobosa* (among others).

data show increased levels of AABW in the Vema Channel during glacial-interglacial transitions.

Production of AABW during both glacial and interglacial intervals contrasts with NADW production, which seems to show a clear decrease during interglacial times. The dissimilar records hence suggest no direct link between AABW and NADW circulation. In fact, based on the available data, it seems that changes in oceanographic conditions in the Southern Ocean had little effect on AABW formation.

Reconstruction of sea-ice conditions has proven to be more difficult than the reconstruction of sea-surface temperatures. Studies by Cooke and Hays (1982) and CLIMAP (1981) have suggested that perennial sea-ice cover expanded equatorward by ~5 degrees. Recent work by Burckle and others (1982), however, on the basis of the analysis of diatoms and distribution of sediment type, has suggested that the expanded sea-ice cover was only seasonal. The distinction between seasonal and perennial sea ice is significant in determining the stability of the water column, and thus its susceptibility to overturning and deep-water formation. A perennial ice cover is an indication of a very stable water column which is essentially isolated from any significant buoyancy forcing, due to the insulating nature of sea-ice, much as in the Arctic today. One would thus not expect such a water column to be susceptible to convection. By contrast, a seasonable sea-ice cover is more like the present-day Weddell Sea situation, and deep or bottom water is possible.

An additional complication unique to understanding the past record of AABW production is its many source areas. Presently, it is produced on the continental shelf of the Weddell Sea, in polynyas in the Weddell Sea, beneath the Ross Ice Shelf, and along the continental margins of Antarctica. The relative importance of these sources may have changed in the past. For example, Kellogg (1986) suggested evidence of glacial grounding on the Weddell Sea continental shelf during glacial times, which would eliminate this area as a source region. Moreover, the extent and nature of sea ice would have an important effect on the continental margin areas around Antarctica.

SUMMARY

In this paper, we have presented an overview of modern deep-water circulation and paleoceanographic studies dealing with past circulation conditions. A number of observations can be made based on existing studies.

1. The ecology of benthic foraminifera is more complex than previously thought, due in part to the existence of species-specific microhabitats. Consistent correlations between the abundances of *E. umbonifera* and calcium carbonate undersaturation and *Uvigerina* and percentage of organic carbon/fine-grained sediment make these species useful for paleoenvironmental reconstructions. At this writing, the use of other species for deep-water circulation studies is limited and must await further study of their ecology.

2. Sedimentological indices of deep-water circulation may be influenced by sedimentation changes on a glacial-interglacial cycle independent of circulation conditions. Yet, this method will be important in the future to complement the geochemical tracers because it provides information on the flow properties (velocity, source areas) of deep waters in the past.

3. On the basis of a number of geochemical studies, NADW production decreased during the last glacial interval (Curry and Lohmann, 1982, 1983; Boyle and Keigwin, 1982; Shackleton and others, 1983), or the chemical characteristics of NADW changed (Mix and Fairbanks, 1985). The role of individual sources of glacial NADW is unknown.

4. The presence and nature of sea-ice in the Labrador Sea, southern Norwegian Sea, and Southern Ocean during glacial intervals is problematic due to the lack of sea-ice indices in the geologic record.

5. AABW is produced during both glacial and interglacial intervals, with some variation in production suggested during the glacial to interglacial transitions. The lack of glacial-interglacial cycles suggests that variation of NADW circulation had little effect on AABW production. Sources of AABW during glacial times remain unknown.

6. Changes in Pacific deep-water circulation are suggested on the basis of a variety of data. These changes may have resulted from glacial production of North Pacific deep water or from an increased flux of Southern Ocean Water.

ACKNOWLEDGMENTS

We thank C. D. Hollister and P. H. Wiebe for encouragement to pursue this multidisciplinary study; W. S. Broecker, L. H. Burckle, W. B. Curry, L. D. Keigwin, Jr., T. B. Kellogg, M. S. McCartney, and D. Schnitker for stimulating discussions; L. H. Burckle, W. B. Curry, T. Janecek, L. D. Keigwin, Jr., T. B. Kellogg, M. S. McCartney, A. McIntyre, W. F. Ruddiman, D. Schnitker, R. C. Thunell, and B. A. Warren for reviewing the manuscript. This research was initiated while the authors were at the Woods Hole Oceanographic Institution; it was supported by a grant from the Woods Hole Oceanographic Institution Center for the Analysis of Marine Systems (CAMS), and by National Science Foundation Grants DPP-8216539 to B. H. Corliss, DPP85-01976 to D. G. Martinson and OCE-831539 to T. Keffer.

REFERENCES CITED

Balsam, W., 1981, Late Quaternary sedimentation in the western North Atlantic: Stratigraphy and paleoceanography: Palaeogeography, Palaeoclimatology, Palaeoecology, v. 35, p. 215–240.

Belanger, P. E., 1982, Paleo-oceanography of the Norwegian Sea during the past 130,000 years: Coccolithophorid and foraminiferal data: Boreas, v. 11, p. 29–36.

Belanger, P. E., Curry, W. B., and Matthews, R. K., 1981, Core-top evaluation of the paleo-oceanographic utility of benthic foraminiferal isotopic ratios: Palaeogeography, Palaeoclimatology, Palaeoecology, v. 38, p. 205–220.

Booth, J. D., and Burckle, L. H., 1976, Displaced Antarctic diatoms in the southwestern and central Pacific: Pacific Geology, v. 11, p. 99–108.

Boyle, E. A., and Keigwin, L. D., 1982, Deep circulation of the North Atlantic over the last 200,000 years: Geochemical evidence: Science, v. 218, p. 784–787.

——— 1985, Comparison of Atlantic and Pacific paleochemical records for the last 215,000 years: Changes in deep ocean circulation and chemical inventories: Earth and Planetary Science Letters, v. 76, p. 135–150.

Bremer, M. L., 1983, Abyssal benthonic foraminifera and the carbonate saturation of water and a benthonic foraminiferal carbonate saturation history for the Cape Verde basin for the last 550,000 years [Ph.D. thesis]: Woods Hole, Massachusetts, Joint Program in Oceanography, Woods Hole Oceanographic Institution–Massachusetts Institute of Technology, 174 p.

Bremer, M. L., and Lohmann, G P., 1982, Evidence for primary control of the distribution of certain Atlantic Ocean benthonic foraminifera by degree of carbonate saturation: Deep-Sea Research, v. 29(8A), p. 987–998.

Brennecke, W., 1921, Die ozeanographischen Arbeiten der Deutschen Antarktischen Expedition 1911-1912: Aus dem archiv der Deutschen Seewarte, Volume 39:1, 216 p.

Broecker, W. S., 1982, Glacial to interglacial changes in ocean chemistry: Progress in Oceanography, v. 11, p. 151–197.

Broecker, W. S., Rooth, C., and Peng, Tsung-Hung, 1985, Ventilation of the deep northeastern Atlantic: Journal of Geophysical Research, v. 90, p. 6940–6944.

Bulfinch, D. L., Ledbetter, M. T., Ellwood, B. B., and Balsam, W. L., 1982, The high-velocity core of the western boundary undercurrent at the base of the U.S. continental rise: Science, v. 215, p. 970–973.

Burckle, L. H., 1981, Displaced Antarctic diatoms in the Amirante Passage: Marine Geology, v. 39, M39–M43.

Burckle, L. H., and Biscaye, P., 1971, Sediment transport by Antarctic Bottom Water through the eastern Rio Grande Rise: Geological Society of America Annual Meeting, p. 518.

Burckle, L. H., and Stanton, D., 1975, Distribution of displaced Antarctic diatoms in the Argentine basin: Beiheft 53 zur Nova Hedwigia, v. 8, p. 283–292.

Burckle, L. H., Abbott, W. H., and Maloney, J., 1973, Sediment transport by Antarctic Bottom Water in the Southeast Indian Ocean: EOS (American Geophysical Union Transactions), v. 54, p. 336.

Burckle, L. H., Venkatarathnam, K., and Booth, J. D., 1974, Sediment transport by Antarctic Bottom Water in the western Indian Ocean: EOS (American Geophysical Union Transactions), v. 55, p. 312.

Burckle, L. H., Robinson, D., and Cooke, G., 1982, Reappraisal of sea-ice distribution in Atlantic and Pacific sectors of the Southern Ocean at 18,000 yr B.P.: Nature, v. 299, p. 435–437.

Burke, S. C., 1981, Recent benthic foraminifera of the Ontong Java Plateau: Journal of Foraminiferal Research, 11(1), p. 1–19.

Clarke, R. A., and Gascard, J. C., 1983, The formation of Labrador Sea Water: Part 1, Large scale processes: Journal of Physical Oceanography, v. 13, p. 1764–1778.

CLIMAP Project Members, 1981, Seasonal reconstruction of the Earth's surface at the last glacial maximum: Geological Society of America Map and Chart Series MC-36.

Cooke, D. W., and Hays, J. D., 1982, Estimates of Antarctic Ocean seasonal sea-ice cover during glacial intervals, *in* Craddock, C., ed., Antarctic Geoscience: Madison, Wisconsin, University of Wisconsin Press, p. 1017–1025.

Corliss, B. H., 1979a, Recent deep-sea benthonic foraminiferal distributions in the southeast Indian Ocean: Inferred bottom-water routes and ecological implications: Marine Geology, v. 31, p. 115–138.

——— 1979b, Quaternary Antarctic bottom-water history: Deep-sea benthonic foraminiferal evidence from the Southeast Indian Ocean: Quaternary Research, v. 12, p. 271–289.

——— 1982, Linkage of North Atlantic and Southern Ocean deep-water circulation during glacial intervals: Nature, v. 298, p. 458–460.

——— 1983a, Distribution of Holocene deep-sea benthonic foraminifera in the Southwest Indian Ocean: Deep-Sea Research, v. 30(2A), p. 95–117.

——— 1983b, Quaternary circulation of the Antarctic circumpolar current: Deep-Sea Research, v. 30(1), p. 47–61.

——— 1985, Microhabitats of benthic foraminifera within deep-sea sediments: Nature, v. 314, p. 435–438.

Curry, W. B., and Lohmann, G. P., 1982, Carbon isotopic changes in benthic foraminifera from the western South Atlantic: Reconstruction of glacial abyssal circulation patterns: Quaternary Research, v. 18, p. 218–235.

——— 1983, Reduced advection into Atlantic Ocean deep eastern basins during last glaciation maximum: Nature, v. 306, p. 577–580.

Damuth, J. E., Jacobi, R. D., and Hayes, D. E., 1983, Sedimentation processes in the northwest Pacific basin revealed by echo-character mapping studies: Geological Society of America Bulletin, v. 94, p. 381–395.

Douglas, R. G., 1981, Paleoecology of continental margin basins: A modern case history from the borderland of southern California, *in* Depositional systems of active continental margin basins: Society of Economic Paleontologists and Mineralogists, Pacific Section, Short Course Notes, p. 121–156.

Douglas, R. G., and Woodruff, F., 1981, Deep-sea benthic foraminifera, *in* Emiliani, C., ed., The oceanic lithosphere, the sea: New York, Wiley-Interscience, Volume 7, p. 1233–1327.

Duplessy, J. C., 1978, Isotope studies, *in* Gribbin, J., ed., Climatic change: London, England, Cambridge University Press, p. 46–67.

Duplessy, J. C., and Shackleton, N. J., 1985, Response of global deep-water circulation to Earth's climatic change 135,000–107,000 years ago: Nature, v. 316, p. 500–507.

Duplessy, J. C., Chenouard, L., and Vila, F., 1975, Weyl's theory of glaciation supported by isotopic study of Norwegian Core K11: Science, v. 188, p. 1208–1209.

Duplessy, J. C., Moyes, J., and Pujol, C., 1980, Deep water formation in the North Atlantic Ocean during the last ice age: Nature, v. 286, p. 479–482.

Ellwood, B. B., 1980, Application of the anisotropy of magnetic susceptibility method as an indicator of bottom-water flow direction: Marine Geology, v. 34, M83–M90.

Ellwood, B. B., and Ledbetter, M. T., 1977, Antarctic Bottom Water fluctuations in the Vema Channel: Effects of velocity changes in particle alignment and size: Earth and Planetary Science Letters, v. 35, p. 189–198.

Ellwood, B. B., Ledbetter, M. T., and Johnson, D. A., 1979, Sedimentary fabric: A tool to delineate a high-velocity zone within a deep western Indian Ocean bottom current: Marine Geology, v. 33, p. M51–M55.

Emerson, S., 1985, Organic carbon preservation in marine sediments, *in* Sundquist, E. T., and Broecker, W. S., eds., The carbon cycle and atmospheric CO_2: Natural variations Archean to present: Washington, D.C.: American Geophysical Union Geophysical Monograph 32, p. 78–87.

Fillon, R. H., and Duplessy, J. C., 1980, Labrador Sea bio-, tephro-, oxygen isotopic stratigraphy and late Quaternary paleoceanographic trends: Canadian Journal of Earth Sciences, v. 17, p. 831–854.

Fillon, R. H., and Harmes, R. A., 1982, Northern Labrador shelf glacial chronology and depositional environments: Canadian Journal of Earth Sciences, v. 19, p. 162–192.

Foster, T. D., and Carmack, E. C., 1976, Frontal zone mixing and Antarctic Bottom Water formation in the southern Weddell Sea: Deep-Sea Research, v. 23, p. 301–317.

Georgi, D. T., 1981, Circulation of bottom waters in the southwestern South Atlantic: Deep-Sea Research, v. 28A, p. 959–979.

Gill, A. E., 1973, Circulation and bottom water production in the Weddell Sea: Deep-Sea Research, v. 20, p. 111–140.

Gordon, A. L., 1972a, Spreading of Antarctic Bottom Waters II, *in* Gordon, A. L., ed., Studies in physical oceanography: London, England, Gordon and Breach, p. 1–17.

——— 1972b, Introduction: Physical oceanography of the Southeast Indian Ocean, *in* Hayes, D. E., ed., Antarctic oceanology II: The Australian–New Zealand sector: Antarctic Research Series, v. 19, p. 3–9.

——— 1974, Varieties and variability of Antarctic bottom water: Collagues Internationaux du CNRS, v. 215, p. 33–47.

——— 1978, Deep Antarctic convection west of Maud Rise: Journal of Physical Oceanography, v. 8, p. 600–612.

——— 1982, Weddell Deep Water variability: Journal of Marine Research, v. 40 (suppl.), p. 199–217.

——— 1983, World ocean watermasses and saltiness of the Atlantic, *in* Large scale oceanographic experiments in the World Climate Research Programme: WCRP Publication Series Number 1, World Meteorological Organization, p. 97–108.

Gordon, A. L., and Tchernia, P., 1972, Waters of the continental margin off Adelie Coast, Antarctica, *in* Hayes, D. E., ed., Antarctic oceanology II: The Australian–New Zealand sector: Washington, D.C., American Geophysical Union Antarctic Research Series, v. 19, p. 59–60.

Graham, D. W., Corliss, B. H., Bender, M. L., and Keigwin, L. D., 1981, Carbon and oxygen isotopic disequilibrium of Recent deep-sea benthic foraminifera: Marine Micropaleontology, v. 6, p. 483–497.

Greiner, G.O.G., 1974, Environmental factors controlling the distribution of Recent benthonic foraminifera: Breviora, v. 420, p. 1–35.

Hays, J. D., Imbrie, J., and Shackleton, N. J., 1976, Variations in the Earth's orbit: Pacemaker of the ice ages: Science, v. 194, p. 1121–1132.

Hester, K., and Boyle, E., 1982, Water chemistry control of cadmium content in Recent benthic foraminifera: Nature, v. 298, p. 260–262.

Hobart, M. A., Bunce, E. T., and Sclater, J. G., 1975, Bottom water flow through the Kane Gap, Sierra Leone Rise, Atlantic Ocean: Journal of Geophysical Research, v. 80, p. 5083–5088.

Hogg, N., Biscaye, P., Gardner, W., and Schmitz, W. J., Jr., 1982, On the transport and modification of Antarctic Bottom Water in the Vema Channel: Journal of Marine Research, v. 40 (suppl.), p. 231–263.

Imbrie, J., and Kipp, N. G., 1971, A new micropaleontological method for quantitative paleo-climatology: Application to a late Pleistocene Caribbean core, *in* Turekian, K., ed., The late Cenozoic glacial ages: New Haven, Connecticut, Yale University Press, p. 71–181.

Ivers, W. D., 1975, The deep circulation in the northern North Atlantic Ocean, with especial reference to the Labrador Sea [Ph.D. thesis]: San Diego, California, University of California, 179 p.

Jacobs, S. S., and Georgi, D. T., 1979, Observations on the Southwest Indian/Antarctic Ocean, *in* Angel, M., ed., Voyage of discovery: New York, Pergaman Press, p. 43–84.

Jacobs, S. S., Amos, A. F., and Bruchhausen, P. M., 1970, Ross Sea oceanography and Antarctic Bottom Water formation: Deep-Sea Research, v. 17(5), p. 935–962.

Jacobs, S. S., Gordon, A. L., and Ardai, J. L., Jr., 1979, Circulation and melting beneath the Ross Ice Shelf: Science, v. 203, p. 439–442.

Jansen, E., and Bjorklund, K. R., 1985, Surface ocean circulation in the Norwegian Sea 15,000 B.P. to present: Boreas, v. 14(3), p. 243–257.

Jansen, E., Sejrup, H. P., Fjaeran, T., Hald, M., Holtedahl, H., and Skarbo, O., 1983, Late Weichselian paleoceanography of the southeastern Norwegian Sea: Norsk Geologisk Tiddskrift, v. 63, p. 117–146.

Jenkins, W. J., and Rhines, P. B., 1980, Tritium in the deep North Atlantic Ocean: Nature, v. 286, p. 877–880.

Johnson, D. A., 1984, Research on the Vema Channel: Marine Geology, v. 58, p. 1–274.

Johnson, D. A., Ledbetter, M. T., and Burckle, L. H., 1977, Vema Channel paleoceanography: Pleistocene dissolution cycles and episodic bottom water flow: Marine Geology, v. 23, p. 1–33.

Jones, G., and Johnson, D. A., 1984, Displaced Antarctic diatoms in Vema Channel sediments: Late Pleistocene/Holocene fluctuations in AABW flow: Marine Geology, v. 58, p. 165–186.

Kanaev, V. F., and others, 1977, Indiiski: Okean: Academy of Sciences of the U.S.S.R., O.V. Schmidt Institute of the Physics of the Earth and P. P. Shirshov Institute of Oceanology. Central Board of Geodesy and Cartography under the Council of Ministers of the U.S.S.R., Moscow, 4 sheets, scale 1:10,000,000.

Keffer, T., 1985, The ventilation of the world's oceans: Maps of the potential vorticity field: Journal of Physical Oceanography, v. 15, p. 509–523.

Kellogg, T. B., 1975, Late Quaternary climatic changes in the Norwegian and Greenland Seas, *in* Weller, G., and Bowling, S. A., eds., Climate of the Arctic: Fairbanks, Alaska, University of Alaska, p. 3–36.

——— 1976, Late Quaternary climatic changes: Evidence from deep-sea cores of Norwegian and Greenland Seas, *in* Cline, R. M., and Hays, J. D., eds., Investigation of late Quaternary paleoceanography and paleoclimatology: Geological Society of America Memoir 145, p. 77–110.

——— 1977, Paleoclimatology and paleo-oceanography of the Norwegian and Greenland Seas: The last 450,000 years: Marine Micropaleontology, v. 2, p. 235–249.

——— 1980, Paleoclimatology and paleo-oceanography of the Norwegian and Greenland Seas: Glacial-interglacial contrasts: Boreas, v. 9, p. 115–137.

——— 1984, Paleoclimatic significance of subpolar foraminifera in high-latitude marine sediments: Canadian Journal of Earth Sciences, v. 21, p. 189–193.

——— 1985, Comment on 'Early deglaciation of the Greenland Sea during the last glacial to interglacial transition': Marine Geology, v. 62, p. 167–173.

——— 1986, Glacial-interglacial changes in global deep-water characteristics: Paleoceanography (in press).

Kellogg, T. B., Duplessy, J. C., and Shackleton, N. J., 1978, Planktonic foraminiferal and oxygen isotopic stratigraphy and paleoclimatology of Norwegian Sea deep-sea cores: Boreas, v. 7, p. 61–73.

Killworth, P. D., 1977, Mixing on the Weddell Sea continental slope: Deep-Sea Research, v. 24, p. 427–448.

Ledbetter, M. T., 1979, Fluctuations of Antarctic Bottom Water velocity in the Vema Channel during the last 160,000 years: Marine Geology, v. 33, p. 71–89.

——— 1984, Bottom-current speed in the Vema Channel during the late Quaternary based on particle-size analyses: Marine Geology, v. 58, p. 137–149.

Ledbetter, M. T., and Balsam, W. L., 1985, Paleoceanography of the Deep Western Boundary Undercurrent on the North American continental margin for the past 25,000 yr: Geology, v. 13, p. 181–184.

Ledbetter, M. T., and Ellwood, B. B., 1980, Spatial and temporal changes in bottom-water velocity and direction from analysis of particle size and alignment in deep-sea sediments: Marine Geology, v. 38, p. 245–261.

——— 1982, Variations in particle alignment and size in sediments of the Vema Channel record Antarctic Bottom-Water velocity changes during the last 400,000 years, *in* Craddock, C., ed., Antarctic geoscience: Madison, Wisconsin, The University of Wisconsin Press, p. 1033–1038.

Ledbetter, M. T., and Johnson, D. A., 1976, Increased transport of Antarctic bottom water in the Vema Channel during the last ice age: Science, v. 194, p. 837–839.

Lohmann, G. P., 1978a, Abyssal benthonic foraminifera as hydrographic indicators in the western South Atlantic Ocean: Journal of Foraminiferal Research, v. 8(1), p. 6–34.

——— 1978b, Response of the deep-sea to ice ages: Oceanus, p. 58–64.

Lutze, G. F., 1980, Depth distribution of benthonic foraminifera on the continental margin off NW Africa: Meteor Forsch. Erg C, v. 33, p. 31–80.

Lutze, G. F., and Coulbourn, W. T., 1984, Recent benthic foraminifera from the continental margin of northwest Africa: Community structure and distribution: Marine Micropaleontology, v. 8, p. 361–401.

Mammerickx, J., 1985, A deep-sea thermohaline flow path in the northwest Pacific: Marine Geology, v. 65, p. 1–19.

Mangini, A., Dominik, J., Muller, P. J., and Stoffers, P., 1982, Pacific deep circulation: A velocity increase at the end of the interglacial stage 5?: Deep-Sea Research, v. 29(12A), p. 1517–1530.

Mantyla, A. W., 1975, On the potential temperature in the abyssal Pacific Ocean: Journal of Marine Research, v. 33, p. 341–354.

Martinson, D. G., Killworth, P. D., and Gordon, A., 1981, A convective model for the Weddell Polynia: Journal of Physical Oceanography, v. 11, p. 466–488.

McCartney, M. S., and Talley, L. D., 1982, The subpolar mode water of the North Atlantic Ocean: Journal of Physical Oceanography v. 12, p. 1169–1188.

——— 1984, Warm-to-cold water conversions in the northern North Atlantic

Ocean: Journal of Physical Oceanography, v. 14, p. 922–935.
McCartney, M. S., Bennett, S. L., Raymer, M. E., and Bremer, M. L., 1986, Eastward flow through the Mid-Atlantic Ridge fracture at 11°N and its influence on the abyssal water mass of the eastern basin: Journal of Physical Oceanography (in press).
Metcalf, W. G., Heezen, B. C., and Stalcup, M. C., 1964, The sill depth of the Mid-Atlantic Ridge in the equatorial region: Deep-Sea Research, v. 11, p. 1–10.
Miller, K. G., and Lohmann, G. P., 1982, Environmental distribution of Recent benthic foraminifera on the northeast United States continental slope: Geological Society of America Bulletin, v. 93, p. 200–206.
Mix, A. C., and Fairbanks, R. G., 1985, North Atlantic surface-ocean control of Pleistocene deep-ocean circulation: Earth and Planetary Science Letters, v. 73, p. 231–243.
Moore, T. C., Jr., Burckle, L. H., Geitzenauer, K., Luz, B., Molina-Cruz, A., Robertson, J. H., Sacks, H., Sancetta, C., Thiede, J., Thompson, P., and Wenkam, C., 1980, The reconstruction of sea surface temperatures in the Pacific Ocean of 18,000 B.P.: Marine Micropaleontology, v. 5, p. 215–247.
Mosby, H., 1934, The waters of the Atlantic Antarctic Ocean, *in* Scientific results of the Norwegian Antarctic Expedition 1927–1928, Volume 1(11), 131 p.
Mudie, P. J., and Aksu, A. E., 1984, Palaeoclimate of Baffin Bay from 300,000 year record of foraminifera, dinoflagellates and pollen: Nature, v. 312, p. 630–634.
Muller, P. J., Erlenkeuser, H., and Von Grafenstein, R., 1983, Glacial-interglacial cycles in oceanic productivity inferred from organic contents in eastern North Atlantic sediment cores, *in* Thiede, J., and Suess, E., eds., Coastal upwelling—Its sediment record, Part B: Sedimentary records of ancient coastal upwelling: New York, Plenum Press.
Murray, J. W., 1973, Distribution and ecology of living benthic foraminiferids: London, England, Heinemann Educational Books, 277 p.
Nansen, F., 1902, The oceanography of the North Polar Basin, *in* Nansen, F., ed., The Norwegian North Polar Expedition, 1893–1896; Scientific results: New York, Longmans, Green and Co., v. 3(9), 427 p.
Olausson, E., 1981, On the glacial Norwegian-Greenland Seas and the Arctic Ocean: Quaternary Studies in Poland, v. 3, p. 79–89.
Peterson, L. C., 1984, Recent abyssal benthic foraminiferal biofacies of the eastern Indian Ocean: Marine Micropaleontology, v. 8, p. 479–519.
Peterson, W. H., and Rooth, C.G.H., 1976, Formation and exchange of deep water in the Greenland and Norwegian Seas: Deep-Sea Research, v. 23, p. 273–283.
Reid, J. L., 1979, On the contribution of the Mediterranean outflow to the Norwegian-Greenland Sea: Deep-Sea Research, v. 26, p. 1199–1223.
Reid, J. L., Nowlin, W. D., Jr., and Patzert, W. C., 1977, On the characteristics and circulation of the southwestern Atlantic Ocean: Journal of Physical Oceanography, v. 7, p. 62–91.
Rodman, M. R., and Gordon, A. L., 1982, Southern Ocean bottom water of the Australian-New Zealand sector: Journal of Geophysical Research, v. 87, p. 5771–5778.
Sancetta, C., 1979, Oceanography of the North Pacific during the last 18,000 years: Evidence from fossil diatoms: Marine Micropaleontology, v. 4, p. 103–123.
Saunders, P. M., 1982, Circulation in the eastern North Atlantic: Journal of Marine Research, v. 40 (suppl.), p. 641–657.
Schnitker, D., 1974, West Atlantic abyssal circulation during the past 120,000 years: Nature, v. 248 (5447), p. 385–387.
——— 1979, The deep waters of the western North Atlantic during the past 24,000 yrs, and the re-initiation of the western boundary current: Marine Micropaleontology, v. 4, p. 265–280.
Sejrup, H. P., Jansen, E., Erlenkeuser, H., and Holtedahl, H., 1984, New faunal and isotopic evidence on the late Weichselian–Holocene oceanographic changes in the Norwegian Seas: Quaternary Research, v. 21, p. 74–84.
Shackleton, N. J., 1977, Carbon-13 in Uvigerina: Tropical rain forest history and the equatorial Pacific carbonate dissolution cycles, *in* Anderson, N. R., and Malahoff, A., eds., The fate of fossil fuel CO_2 in the oceans: New York, Plenum Publishing Co.
Shackleton, N. J., and Duplessy, J. C., 1985, Carbon isotope evidence for formation of North Pacific deep water at 18 ka BP glacial maximum: EOS (American Geophysical Union Transactions), v. 66, p. 292.
Shackleton, N. J., and Opdyke, N. D., 1973, Oxygen isotope and paleomagnetic stratigraphy of equatorial Pacific core V28-238: Oxygen isotope temperatures and ice volumes on a 10^5 year and 10^6 year scale: Quaternary Research, v. 3, p. 39–55.
Shackleton, N. J., Imbrie, J., and Hall, M. A., 1983, Oxygen and carbon isotope record of East Pacific core V19-30: Implications for the formation of deep water in the late Pleistocene North Atlantic: Earth and Planetary Science Letters, v. 65, p. 233–244.
Streeter, S. S., 1973, Bottom water and benthonic foraminifera in the North Atlantic—Glacial-interglacial contrasts: Quaternary Research, v. 3, p. 131–141.
Streeter, S. S., and Lavery, S. A., 1982, Holocene and latest glacial benthic foraminifera from the slope and rise off eastern North America: Geological Society of America Bulletin, v. 93, p. 190–199.
Streeter, S. S., and Shackleton, N. J., 1979, Paleocirculation of the deep North Atlantic: 150,000-year record of benthic foraminifera and oxygen-18: Science, v. 203, p. 168–171.
Streeter, S. S., Belanger, P. E., Kellogg, T. B., and Duplessy, J. C., 1982, Late Pleistocene paleo-oceanography of the Norwegian-Greenland Sea: Benthic foraminiferal evidence: Quaternary Research, v. 18, p. 72–90.
Swift, J. H., Aagaard, K., and Malmberg, S.-A., 1980, The contribution of the Denmark Strait overflow to the deep North Atlantic: Deep-Sea Research, v. 27A, p. 29–42.
Talley, L. D., and McCartney, M. S., 1982, Distribution and circulation of Labrador Sea Water: Journal of Physical Oceanography, v. 12, p. 1189–1205.
Tappa, E., and Thunell, R., 1984, Late Pleistocene glacial/interglacial changes in planktonic foraminiferal biofacies and carbonate dissolution patterns in the Vema Channel: Marine Geology, v. 58, p. 101–122.
Theyer, F., 1971, Size-depth variation in *Cyclammina cancellata* Brady, Peru-Chile Trench area: Antarctic Research Series, v. 15, p. 309–313.
Thompson, P. R., 1981, Planktonic foraminifera in the western North Pacific during the past 150,000 years: Comparison of modern and fossil assemblages: Palaeogeography, Palaeoclimatology, Palaeoecology, v. 35, p. 241–279.
Thompson, P. R., and Shackleton, N. J., 1980, North Pacific palaeoceanography: Late Quaternary coiling variations of planktonic foraminifer *Neogloboquadrina pachyderma*: Nature, v. 287, p. 829–833.
Vilks, G., and Mudie, P. J., 1978, Early deglaciation of the Labrador Shelf: Science, v. 202, p. 1181–1183.
Walsh, J. J., 1966, The vertical distribution of phytoplankton in the waters around the Palmer Peninsula [M.S. thesis]: Miami, Florida, University of Miami, p. 1–162.
Warren, B. A., 1977, Deep western boundary current in the eastern Indian Ocean: Science, v. 196, p. 53–54.
——— 1981a, Deep circulation of the world ocean, *in* Warren, B. A., and Wunsch, C., eds., Evolution of physical oceanography: Cambridge, Massachusetts, The M.I.T. Press, p. 6–41.
——— 1981b, Transindian hydrographic section at lat. 18S: Property distributions and circulation in the South Indian Ocean: Deep-Sea Research, v. 28A, p. 759–788.
——— 1983, Why is no deep water formed in the North Pacific?: Journal of Marine Research, v. 41, p. 327–347.
Warren, B. A., Stommel, H., and Swallow, J. C., 1966, Water masses and patterns of flow in the Somali basin during the southwest monsoon of 1964: Deep-Sea Research, v. 13, p. 825–860.
Weyl, P. K., 1968, The role of the oceans in climatic change: A theory of the Ice Ages, *in* Causes of climatic change: Meteorological Monographs, v. 8.
Whitehead, J. A., and Worthington, L. V., 1982, The flux and mixing rates of Antarctic Bottom Water within the North Atlantic: Journal of Geophysical Research, v. 87, C10, p. 7903–7924.
Worthington, L. V., 1970, The Norwegian Sea as a Mediterranean basin: Deep-Sea Research, v. 17, p. 77–84.
——— 1976, On the North Atlantic circulation: Johns Hopkins Oceanographic Studies, Volume 6: Baltimore, Johns Hopkins Press.
Worthington, L. B., and Wright, W. R., 1970, North Atlantic Ocean atlas: Woods Hole, Massachusetts, Woods Hole Oceanographic Institution.
Wright, W. R., and Worthington, L. B., 1970, The water masses of the North Atlantic Ocean; A volumetric census of temperature and salinity: Serial Atlas of the Marine Environment, Folio 19: New York, American Geographical Society, 8 p. and 7 pls.
Zwally, H. J., Wilheit, T. T., Gloersen, P., and Mueller, J. L., 1976, Characteristics of Antarctic sea ice as determined by satellite-borne microwave imagers, *in* Proceedings from the Symposium on Meteorological Observations from Space: Their Contributions to the First GARP Experiment: Boulder, Colorado, p. 94–97.

MANUSCRIPT RECEIVED BY THE SOCIETY SEPTEMBER 23, 1985
REVISED MANUSCRIPT RECEIVED APRIL 9, 1986
MANUSCRIPT ACCEPTED APRIL 21, 1986
WOODS HOLE OCEANOGRAPHIC INSTITUTION CONTRIBUTION NO. 6117
LAMONT-DOHERT GEOLOGICAL OBSERVATORY CONTRIBUTION NO. 3996